GEOLOGY OF THE NETHERLANDS

Geology of the Netherlands

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Aerial view of ‘Het Verdronken Land van Saeftinghe’ (the drowned land of Saeftinghe), presently a tidal area situated along the Westerschelde estuary in the province of Zeeland. The area measures ca. 30 km^2 and was inhabited until the end of the 16^{th} century.
Courtesy of the Ministry of Transport and Public Works (Ministerie van Verkeer en Waterstaat).

Contents

NAM and geology: a long-standing and deep relationship

During the 1930s many people thought that the Netherlands had few underground resources. There were only coal mines in the southern Dutch province of Limburg and rock salt (halite) mines in Boekelo, Twente. Starting in 1943, this picture changed dramatically. The riches lying beneath the Netherlands were discovered.

In that year, the Shell subsidiary Exploratie Nederland discovered an oil field near Schoonebeek in the eastern Netherlands. After this discovery, Shell and Esso decided to jointly invest capital in a new company that would search for and extract oil: the *Nederlandse Aardolie Maatschappij* (Dutch Oil Company), or NAM. NAM was founded on 19 September 1947.

In 1948, NAM made its first discovery of natural gas in Coevorden. A little more than ten years later (1959), NAM uncovered further secrets of the Earth's crust. Near Slochteren in the northern Netherlands, the company drilled into the famous Groningen gas field. This field is one of the world's largest natural gas fields. It originally held 2.7 trillion (2.7×10^{12}) cubic metres of extractable natural gas. This find opened the door to the exploration for and production of natural gas under the North Sea. In 1961, NAM was the first Western European company to drill for gas in the North Sea.

Today, NAM is the Netherlands' largest producer of natural gas. Per year, NAM produces around 50 billion cubic metres of gas. A little more than half of this gas comes from the Groningen gas field and the rest comes from various smaller fields elsewhere on land or in the North Sea. NAM produces around 75% of the total Dutch gas production. Oil is also still being extracted by NAM, although in much smaller quantities than the gas produced. NAM is responsible for around 25% of the Netherlands' oil production.

By using the Earth's crust for storing natural gas, NAM is anticipating future needs. Examples are the underground gas storage facilities in Langelo and Grijpskerk (both in the northern Netherlands), which allow NAM to continue to meet the demand for gas during periods of extreme cold. These gas storage facilities are in empty gas fields. Additionally, NAM is still searching for new gas fields under the Netherlands and the North Sea. On the basis of the gas reserves that have been verified already, NAM estimates that the Netherlands has sufficient gas reserves for another 25 years, depending, of course, on the demand for natural gas during the coming years.

Right from the beginning, NAM has always had an intensive relationship with geologists. This is no different today. NAM still wants to uncover the secrets of the Earth's crust. NAM and geology – a long and deep relationship.

Foreword

The underground mountains of the Netherlands

Somehow the Netherlands has always managed to keep away from the orogenic turmoil that shaped most of Europe. It kept itself in the wake of the Caledonian mountain front, it barely escaped the Hercynian orogeny that shaped the Ardennes in the south, it kept its integrity during the break-up of Pangaea apart from minor Jurassic outpourings, and only received distant echos of the Cenozoic collision of Africa with Europe which gave rise to the Alps. It contented itself with receiving the debris of the newly formed mountain belts in its ever shifting Carboniferous, Permo-Triassic and Cenozoic deltas and fans during an almost uninterrupted history of tectonic subsidence. It is not a coincidence that the Netherlands is often called Europe's sink. Yet, in its subsurface, it bears the traces of humid tropical swamps, scorching deserts, glistening salt lakes, lukewarm seas and solid ice caps, and made its way across the globe from the location of its former colony Suriname to its present position. Changing climates and changing sea levels provided it with abundant energy resources including peat, coal, oil and natural gas.

The casual observer of the present-day landscape of the Netherlands is largely unaware of this rich history, as unfortunately we have not yet got the means to look as easily through the earth as we look to the stars much farther away. It has taken painstaking mapping, seismic survey, drilling hundreds of thousands of cores during all of the two centuries geology exists as a science, to unravel this picture. And the picture is surprisingly complex. In spite of its proven ability to eschew orogeny, the Netherlands is far from a layer cake. Extensive faulting, both extensional and compressional, made the subsurface a complicated chequerboard. Faulting has set the scene for strong local variations in sedimentation and erosion, and has put formations of widely differing ages in contact with each other. A quick glance at the cross-section of the Netherlands shows the wisdom of saying that our country has its mountains underground.

The editors of this book, Theo Wong, Dick Batjes and Jan de Jager are to be congratulated for bringing together such an excellent collection of papers on the geology of the Netherlands, patiently peeling off the layers of time one by one. From their modest account of previous studies it appears that there has never been such a comprehensive volume on our subsurface and surface geology like the present one. I am convinced that it will remain a standard reference on Dutch geology for decades to come.

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Introduction

The Netherlands is a densely populated country, situated at around 52° N latitude between the North Sea and the Paleozoic massifs of Brabant, the Ardennes and the Rhine, which lie just across the frontier in Belgium and Germany (Fig. 1). Most of the country's surface is flat and cultivated, and present-day sedimentation and erosion are almost everywhere influenced by man: rivers are contained by dikes; many streams are canalized; swamps, lakes and large parts of an inland sea have been turned into polders; and in many places dikes strengthen the coastline. Without dikes about half of the present Dutch land surface would be flooded. To keep the reclaimed polder areas dry and fit for farming, pumping stations – formerly windmills – extract water continuously for transfer into bordering water bodies. This water extraction has led to oxidation and shrinkage of the shallow soil and thus to subsidence. In extensive areas in the west and north of the country several metres of subsidence has occurred since medieval times. In many places the water in a canal, river or lake can be seen at a level well above that of the land behind the dike.

Over the last few decades the ecological and scenic value of nature reserves and of 'nature areas' (natuurgebieden) in general, such as the Waddenzee north of the Dutch mainland and specific areas within certain polders, have become increasingly appreciated, and much effort is now being made to conserve them.

Most of the surface sediments are of Quaternary age. Tertiary, Mesozoic and even Paleozoic rocks occur only locally at or near the surface along the country's eastern and southern borders. The varied, largely siliciclastic sedimentary succession above metamorphic basement reaches in places more than 10 km in thickness. This succession became intricately structured during several tectonic phases. Late Jurassic to Early Cretaceous rifting and Late Cretaceous to Early Tertiary inversion, together with halokinesis of the Permian Zechstein salt, had the greatest impact.

W.C.H. Staring was the first to write a comprehensive book on the geology of the Netherlands, and he also compiled the first 1 : 200 000 geological map of the country. The two volumes of his *De bodem van Nederland* (The soil of the Netherlands), together exceeding 900 pages, and meagrely illustrated by present-day standards, were published in 1856 and 1860. Volume I deals with the Alluvium (Holocene) and volume II with the Diluvium (Pleistocene) and older rocks. In 1769–1771, J. le Francq van Berkhey had previously published a detailed geological description of 'Holland', i.e. the area of the present-day provinces Noord- and Zuid-Holland. The illustrations of this earlier description notably include a profile, based on shallow drillings and field observations, showing the distribution of sand, clay and peat layers down to 5 m depth over a distance of 130 km (Zagwijn, 2004).

In the beginning of the 20th century, a state-funded exploration campaign for mineral resources, directed by W.A.J.M. van Waterschoot van der Gracht, provided the first insight into the deeper subsurface. This campaign located exploitable deposits of coal and rock salt. Its results were published in detail (Van Waterschoot van der Gracht, 1909, 1918).

An overview in English, entitled *Geological history of the Netherlands*, edited by A.J. Pannekoek and covering both the geology and the geological resources, appeared in 1956, along with a version in Dutch. This overview was published jointly by the Royal Netherlands Geological and Mining Society (KNGMG) and the Geological Foundation – the geological survey at that time – and was intended as an explanation to the Society's 1 : 200 000 geological map, which had been prepared by the Foundation on the basis of its own 1 : 50 000 map. In 1963, *Transactions of the Jubilee Convention* was published, following the 1962 convention on the theme 'Geology and mining in the Netherlands'. Together, the book edited by Pannekoek and the 1963 Transactions give an excellent overview of the state of geological knowledge four to five decades ago. Much of what was presented in that overview has since evolved considerably, mainly as a result of hydrocarbon exploration, both on land and in the North Sea, and the accompanying technological development and scientific progress. For example:

- the Groningen gas field, not immediately appreciated as the giant it is, was discovered in 1959,
- the coal mining in Limburg, once a flourishing industry, ended in 1974, partly as a result of the Groningen discovery,
- the area of the Dutch North Sea sector, blank or barely shown on most maps in the 1963 Transactions, has now been mapped geologically down to several kilometres depth,
- in the process, the name 'Central Graben', originally used for a graben on land, now known as Roer Valley Graben, has come to indicate a much larger graben in the North Sea,
- the effects but not yet the full dynamics of the tectonic inversion, which affected both grabens and also depocentres elsewhere, have become better understood,
- a buried volcano of Late Jurassic age was discovered under the Waddenzee,
- earthquakes induced by the production of natural gas have been not only felt but also analysed and explained.

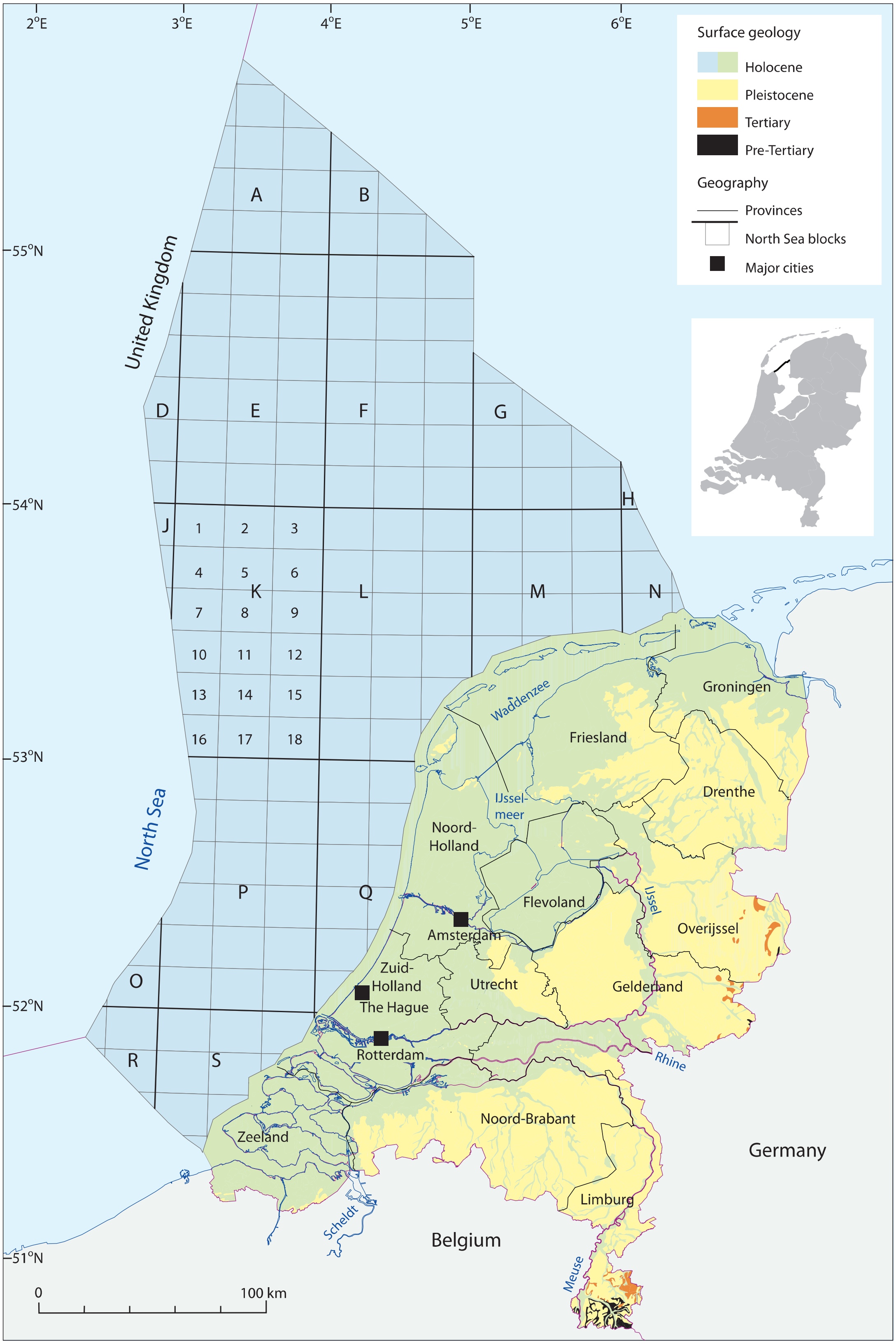

Fig. 1. General map of the geography and geology of the Netherlands (based on TNO data). The Holocene is indicated in blue in the offshore sector outside the territorial waters, and is shown in green elsewhere. The inset map shows the distribution of water and land in the onshore area.

Many more such events or developments have taken place in the period described, and are documented in numerous publications, many of which are in English. A few of these publications merit to be mentioned here.

A set of 1 : 1 000 000 depth and subcrop maps of the entire Netherlands and its directly adjacent offshore area was published by P. Heybroek in 1974, together with an explanatory text. It was a much consulted paper at the time.

In 1979, the Geological Survey published the summary overview *The geology of the Netherlands* (Van Staalduinen et al., 1979).

The book *Seventy-five years of geology and mining in the Netherlands*, published in 1987 by the KNGMG, presents interesting historical reviews of: the geological survey; the mining legislation; the university education in geology and mining engineering; the activities of associations of amateur geologists; and the exploration and exploitation of geological resources (Visser et al., 1987). The book also discusses the further utilization of the subsurface, and it ends with an extensive bibliography.

The stratigraphic nomenclature of the country was compiled from 1993 to 1997 in a well-illustrated, thick volume on behalf of the Geological Survey (RGD) and the Netherlands Oil and Gas Exploration and Production Association (NOGEPA). The volume also contains a summary description of main tectonic elements (Van Adrichem Boogaert & Kouwe, 1993–1997). It updates the nomenclature published in 1980 by the Nederlandse Aardolie Maatschappij and the Geological Survey. For the Quaternary, both nomenclatures refer to the summary compilation which the Survey published in 1979.

A selection of papers on *The geology of gas and oil under the Netherlands* appeared in a KNGMG publication in 1996. The subjects covered range from reservoirs in single oil and gas fields to reviews of the hydrocarbon habitat in the West Netherlands Basin and the regional Rotliegend facies distribution (Rondeel et al., 1996).

In 2004 the Netherlands Institute of Applied Geoscience TNO (NITG), now incorporated in TNO Built Environment and Geosciences, published a bilingual English–Dutch atlas compiling on a 1 : 1 000 000 scale the results of its mapping project on the subsurface geology of the onshore Netherlands. The same institute published in 2003 a book in Dutch entitled *De ondergrond van Nederland* (The subsurface of the Netherlands) on the country's geology and geological resources (De Mulder et al., 2003). Whilst this Introduction was being finalized, Duin et al. (2006) published an article including a set of depth and thickness maps, based on TNO data, covering both the onshore and offshore Netherlands. Several illustrations in the just mentioned book, atlas and article have been included in the present volume. Another TNO book that deserves mention here for those who read Dutch is the historical review by Faasse (2002) entitled *De ontdekking van de ondergrond – Anderhalve eeuw toegepast geowetenschappelijk onderzoek in Nederland* (The discovery of the subsurface – A century and a half of applied geoscientific research in the Netherlands).

Our book begins with an overview of the geological development of the Netherlands, and then follows the stratigraphy upwards in eight chapters from pre-Silesian to Quaternary. The content of the chapters for the deeper subsurface has been somewhat dependent on the availability and incorporation of data that are governed by mining law. According to the new mining law, data acquired after the effective date January 1st, 2003 remain confidential for a period of five years and have to be released subsequently. Data acquired before the effective date become public ten years after their acquisition. The chapters have benefited from some voluntary data release by the companies involved. For reasons of local convenience, some chapters apply a chronostratigraphy that deviates, as indicated in the text, from the nomenclature advocated by the International Commission on Stratigraphy (Gradstein et al., 2004).

Following the chapter 'Quaternary', two chapters deal with magmatism and earthquakes respectively. Five chapters on mineral resources follow, each with its own specific approach, with discussion of: hydrocarbons; peat, coal and coalbed methane; salt; groundwater; and surface minerals. One observation arising from these chapters is that the southern part of the province of Limburg has produced all the coal, lignite and silica sand, as well as most of the limestone and gravel that has been or still is exploited in the Netherlands. The same area was also the site of the country's strongest recorded earthquake and of several finds of large fossil mosasaurs. The final chapters briefly review underground storage and geothermal energy.

The responsibility for the chapter content lies with the authors and not with the institutions or companies to which the authors are or were affiliated.

Inevitably, the book refers to various publications in Dutch. Where both English and Dutch versions are available, e.g. for several works originating in TNO and the former RGD, only the English version has been listed in the bibliography at the end of each chapter.

We owe many thanks to the authors for their chapters and to the reviewers for their comments. We particularly mention the authors who completed their manuscripts at an early stage and who later on updated their contributions. We are much indebted to Bozena van Trigt who compiled the Index; to the Department of Geomedia of the Faculty of Geosciences of Utrecht University who took care of many of the figures; to TNO for their permission to reproduce illustrations; to the Royal Netherlands Academy of Arts and Sciences for undertaking to publish the book;

to Ab van Adrichem Boogaert and Mark Geluk for their reviews of this Introduction; to Guy Smith for adjusting the English of the Introduction's penultimate version; and to Michiel van der Meulen for preparing Figure 1. Finally, we are most grateful to the Nederlandse Aardolie Maatschappij for their generous contribution to the printing costs of this book.

Theo Wong
Dick Batjes
Jan de Jager

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Geological development

J. de Jager

ABSTRACT

The geological evolution of the Netherlands resulted in a highly structured and surprisingly varied subsurface geology below a deceivingly flat topography. In much of the Dutch area, more than 10 km of predominantly siliciclastic sediments overlie the metamorphic basement. These sediments comprise several major unconformities, but on the whole the geological record is almost continuous from the Late Paleozoic onwards. The basins, platforms and highs in the subsurface formed in response to world-wide reorganisations of lithospheric plates. The main tectonic events that affected the area were: i) the Caledonian and Variscan orogenies, resulting in the assembly of the Pangea supercontinent during the Paleozoic, ii) Mesozoic rifting, accompanying the break-up of Pangea, iii) Alpine inversion, resulting from the collision of Africa and Europe during the Late Cretaceous and Early Tertiary, and iv) Oligocene to recent development of the Rhine Graben rift system. Notwithstanding the diversity of these events, a high degree of fault parallelism occurs. The general structural model is, therefore, one of repeated (oblique) reactivation of basement faults which continue to control the structural grain, despite changes in tectonic regime and stress direction. The presence of thick Permian Zechstein salt in much of the subsurface not only resulted in extensive halokinesis, but also caused extensional and transpressional faulting above the salt to be decoupled from sub-salt faulting. The initiation of salt movement resulted in most cases from basement faulting. Overpressures in the Dutch subsurface are caused primarily by burial during the Tertiary. Their preservation is strongly related to effective sealing.

Keywords: Netherlands, rifting, tectonic inversion, halokinesis, Caledonian, Variscan, Saalian, Kimmerian, overpressures

Introduction

In much of the Dutch area, more than 10 km of largely siliciclastic sediments overlie a poorly known metamorphic basement. These sediments comprise several major unconformities, but represent nevertheless an on the whole almost continuous geological record from the Late Paleozoic onwards.

The prominent structural elements in the Dutch subsurface are Late Jurassic to Early Cretaceous extensional and transtensional rift basins, viz. the Dutch Central Graben, the Broad Fourteens Basin, the West and Central Netherlands basins, the Roer Valley Graben and the Lower Saxony Basin (Figs 1, 2). However, over geological time, these elements have continuously changed in response to changing tectonic conditions, as is well illustrated for the onshore area in the Atlas published by the Netherlands Institute of Applied Geoscience TNO (NITG) in 2004, and for the onshore and offshore areas together in a recent paper by Duin et al. (2006). Most of the elements indicated in Figure 1 were active mainly during the Late Jurassic and Early Cretaceous. In this chapter the names of some structural elements are used somewhat loosely; locations are often referred to by the names of the Mesozoic structural elements of Figure 1, rather than by the less common names pertaining to other geological periods. For a formal definition and description of Dutch structural elements reference is made to the Stratigraphic Nomenclature of the Netherlands compiled by Van Adrichem Boogaert & Kouwe (1993–1997).

Many structural events or pulses are recognised in north-west Europe. However, structuration did not occur in discrete events; it was in most cases a more or less continuous process in response to ongoing convergence or divergence of lithospheric plates. Also in modern times lithospheric plates move at a steady pace from or towards each other. Nevertheless, periods of accelerated tectonic activity, punctuating ongoing deformation, can be recognised (Fig. 3). The magmatic activity which accompanied the structural development is reviewed in the chapter by Van Bergen & Sissingh (this volume).

The present chapter consists of four main sections. The first describes the structural history of the Netherlands in the light of plate-tectonic events, and also summarises the overall sedimentary development at group level, the groups being those of Van Adrichem Boogaert & Kouwe (1993–1997). The second section is about the evolution and salient aspects of the main structural elements. The third discusses fault patterns, and the final section deals with the occurrence of overpressures.

Structural history

Paleozoic, Caledonian and Variscan assembly of Pangea

The oldest sediments drilled in the Netherlands are Upper Silurian, folded and densely compacted, fine-grained turbidites in wells on the northern flank of the London-

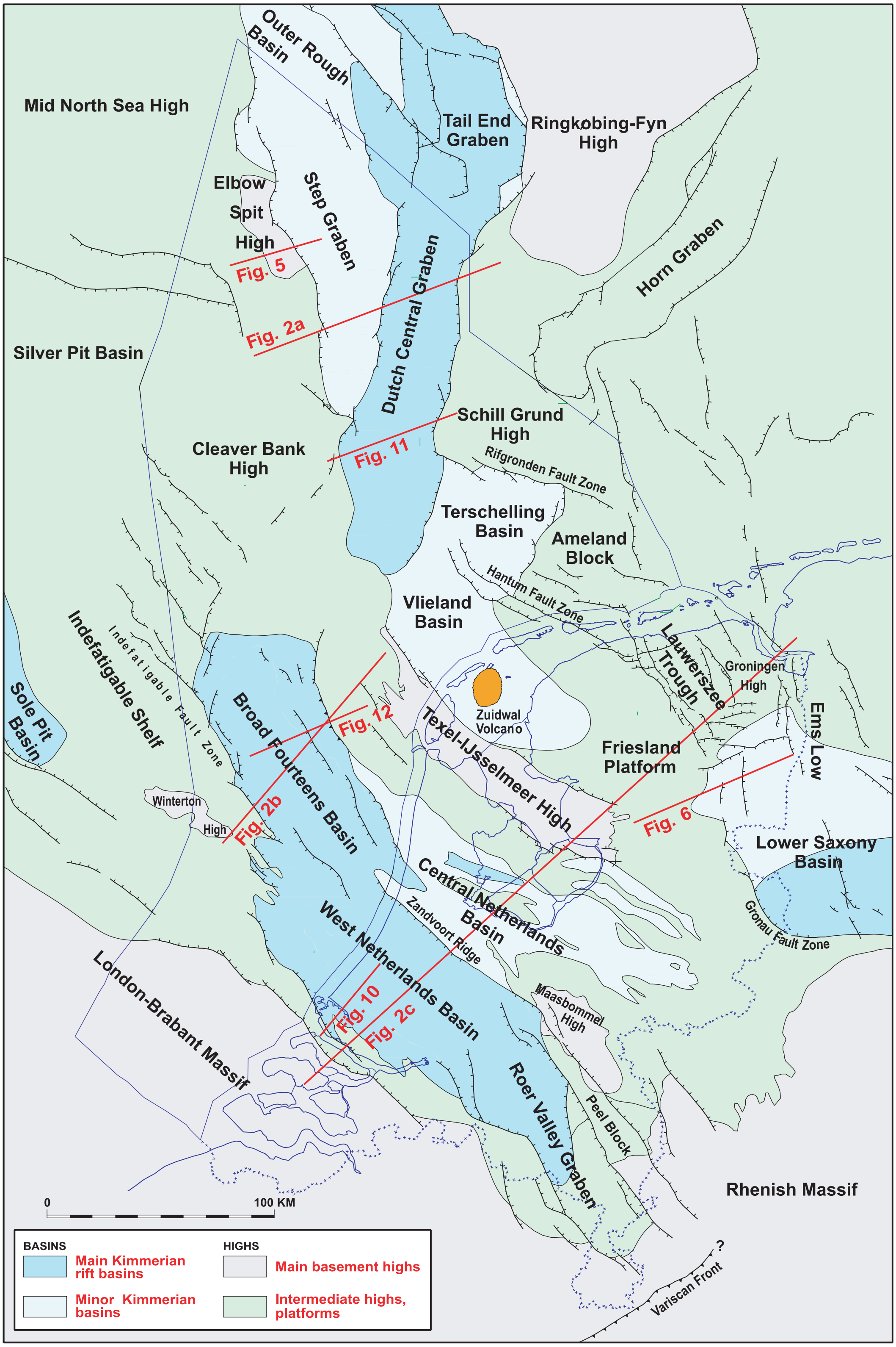


Fig. 1. Structural elements map of the Netherlands showing Mid and Late Kimmerian (Jurassic and Early Cretaceous) basins, highs and platforms. The main basement highs are formed by crystalline basement and Paleozoic, generally without Rotliegend.

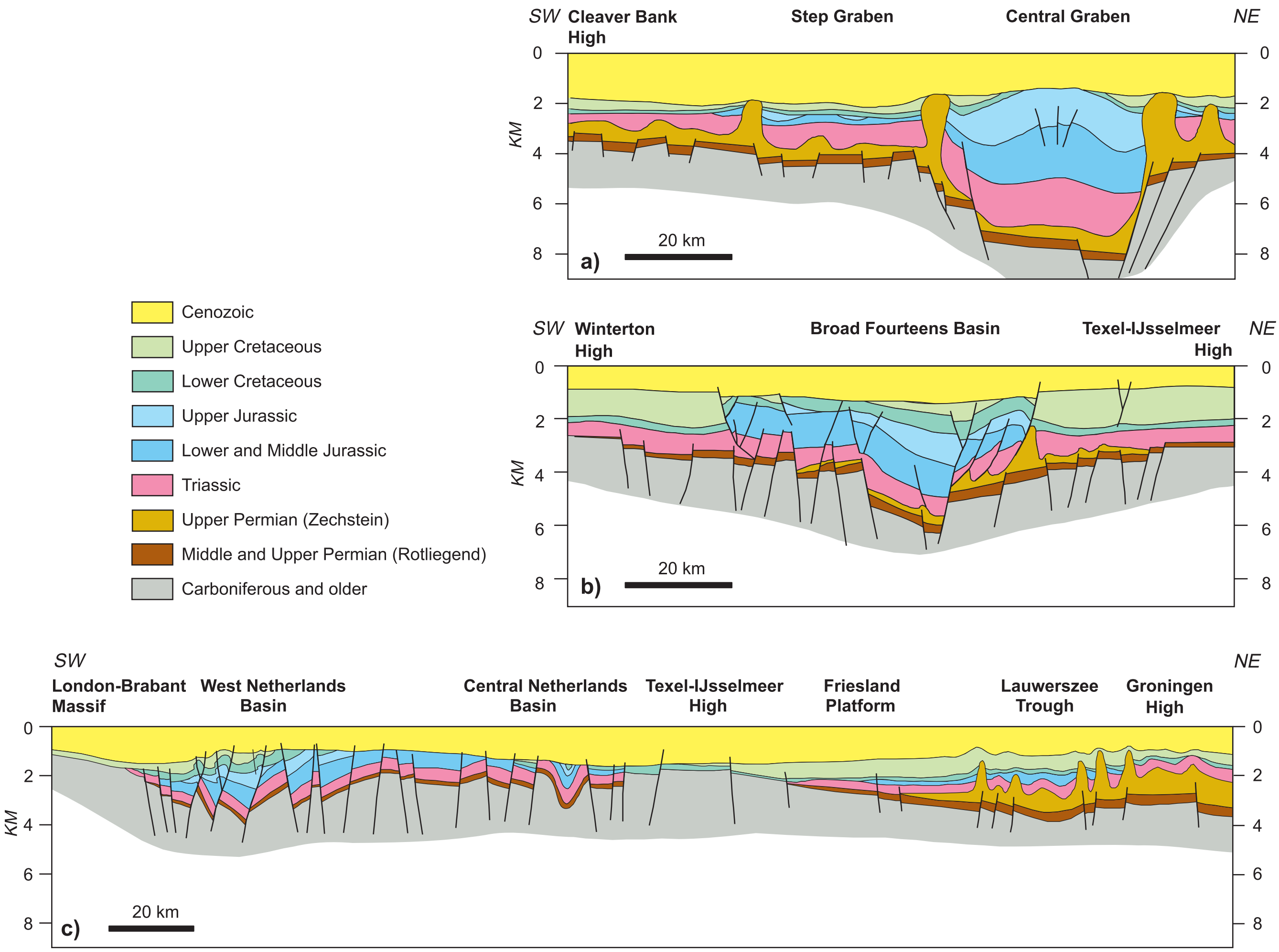


Fig. 2. Regional cross-sections. (a) Dutch northern offshore, (b) Dutch western offshore, (c) Dutch onshore. For locations see Fig. 1.

Brabant Massif. While the period from the Late Silurian to the start of the Permian lasted only half as long as the time from the Permian to the present, deposits laid down during the former period account for probably more than half of the total volume of sediments above the crystalline basement. Because of their deep burial not many wells have penetrated these rocks, in particular the pre-Westphalian deposits (Geluk et al., this volume). Moreover, on seismic reflection data the structure of the older sequences is generally not well imaged. The early structural history of the Netherlands is therefore inferred mainly from regional data. The main authors who have dealt with this are Ziegler (1978, 1981, 1988, 1990a, 1990b), Coward (1993, 1995), Glennie (1986), and Glennie & Underhill (1998), and the following information on the plate-tectonic setting of the Netherlands is based upon their papers. Additional information is given by Geluk et al. (this volume).

The Paleozoic structural history of western Europe was dominated by the convergence of three large continental plates: Laurentia, Baltica and Gondwana. The amalgamation of these plates resulted in the formation of the Pangea supercontinent (Fig. 4). During the Ordovician to Silurian, the Laurentia and Baltica cratons collided to form a single Laurussian continent, with the Caledonian fold belt following the suture of the amalgamated cratons. More or less simultaneously, the continental Gondwana-derived Avalonia terrane, including the London-Brabant Massif, had collided with Baltica to which it was sutured along the North German-Polish Caledonides (Pharaoh et al., 1995). Following the amalgamation of these two larger continents and the smaller terrane, the area of the Netherlands was located close to the triple-junction at some 40 to 30° south of the equator. The Silurian clastics in wells O18-1 and Kortgene-1 appear similar to deposits encountered in East Anglia and in the offshore UK wells 47/29A-1 and 53/16-I, and may have been deposited as distal turbidites in a Caledonian foreland basin (Pharaoh et al., 1995). In northern Belgium, crystalline basement rocks of

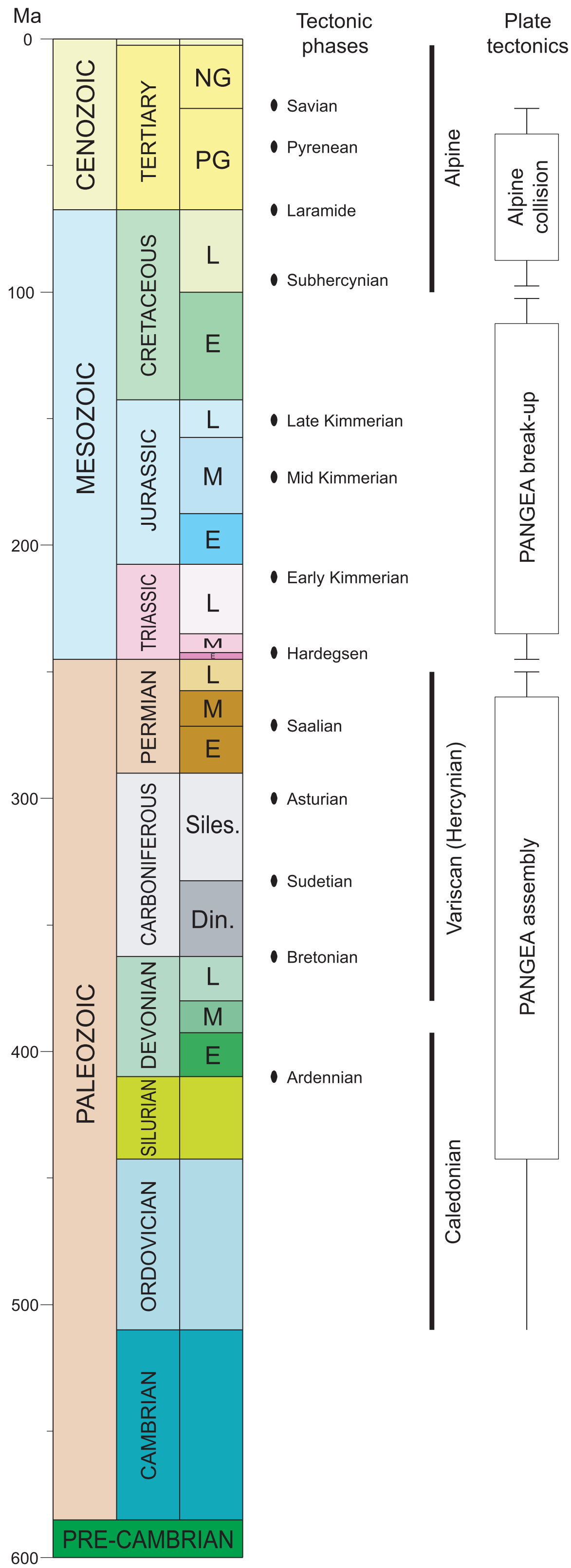


Fig. 3. Tectonic phases in the Netherlands and their relation to plate-tectonic events.

ca. 400 Ma (Late Silurian or Early Devonian) belonging to the London-Brabant Massif are known from outcrop and well data (Ziegler, 1990a). Altered biotiote monzo-granite overlain by non-metamorphosed Devonian Old Red Sandstone on the Elbow Spit High in the northern Dutch offshore (well A17-1, Fig. 5) represents the only direct evidence of Caledonian basement in the subsurface of the Netherlands (Frost et al., 1981; Pharaoh et al., 1995).

Thus, two separate provinces may represent the basement of the Netherlands, namely the Gondwana-derived Avalonia, including the London-Brabant Massif, in the south, and the Caledonian basement in the north. Even though the precise location of the suture between these provinces remains speculative, the general NW-SE trend, that is such a dominant structural feature in the southern half of the Dutch subsurface, may be related to this suture. This would imply a mid-Paleozoic age for this trend.

During the Middle to Late Devonian, when the area of the Netherlands had drifted north to equatorial regions, Gondwana began to converge with Laurussia. During the Visean (Early Carboniferous), full-scale collision resulted in the Himalayan-type Variscan orogeny (Fig. 4). Crustal shortening terminated during the Late Westphalian (Ziegler, 1990a).

While the London-Brabant Massif remained a structural high during the Late Paleozoic, elsewhere sedimentation resumed during the Devonian. Thick carbonate sequences were deposited around the London-Brabant Massif and probably also further north. Deep seismic lines from the Dutch onshore show evidence of large fault blocks bounded by NNW-SSE trending faults that appear to have formed during the Middle Devonian to Early Carboniferous (Fig. 6). The resulting grabens may be related to the NNW-SSE trending Devonian seaway that developed across the eastern Netherlands into what could be referred to as the Proto-Dutch Central Graben (Ziegler, 1988, 1990a).

The Elbow Spit High was a high already during the Devonian. Seismic data clearly show onlapping onto its western flank of Devonian units older than the Devonian red beds of well A17-1 (Fig. 5). Also the Carboniferous units appear to thin onto this high as well as onto the Mid North Sea High. Source-rock maturities measured in Dinantian coals in well E2-1 on the flank of the high suggest that the maximum thickness of the Carboniferous was about 1600 m, i.e. significantly less than elsewhere.

During the Carboniferous, when Gondwana further encroached upon Laurussia, the front of the Variscan fold-and-thrust belt moved north, towards the Netherlands. Its final position runs approximately E-W through Belgium, just south of the Netherlands, swinging to the north-east into Germany (Ziegler, 1990a). At that time, the Netherlands had moved to the northern hemisphere, but still remained in the humid equatorial climate belt.

With the development of the Variscan Mountains, the area of the Netherlands became land-locked. There is no reported evidence of deformation to suggest that significant Variscan compression reached as far north as the Netherlands. The Limburg Group, locally as much as 5500 m thick and consisting of a largely marine and lacustrine Namurian succession and a Westphalian succession of coastal-plain and fluvial-plain deposits with coal seams and thin marine intercalations, accumulated in the foredeep of the fold-and-thrust belt but seems to have been little affected by synsedimentary tectonism. Any faulting seems to have been small-scale, extensional rather than compressional, and diminishing towards the north. At the end of the Westphalian, however, some tectonic movements occurred, heralding the more significant Stephanian and Early Permian, Saalian (in its broader sense) event. In Germany and in the east Netherlands, Stephanian sediments unconformably overlie the Westphalian (Ziegler, 1990a). These youngest Carboniferous sediments are red beds which were laid down when the area of the Netherlands had moved into the arid climate zone of the northern hemisphere at some 15 to 20° latitude.

At the close of the Carboniferous, the area of western Europe became affected by late-Variscan post-orogenic tectonism. Wrench faulting associated with intrusive and extrusive magmatism and thermal uplift caused widespread and deep erosion (Ziegler, 1990a). Broad NW-SE trending swells formed that can be traced on the subcrop map of Westphalian units at the Base Permian Unconformity (Van Buggenum & Den Hartog Jager, this volume). The NW-SE trend that was already established by mid-Paleozoic times was reactivated in response to Early Permian wrench deformation while regionally a conjugate set of NE-SW to NNE-SSW faults developed (Ziegler, 1990a). The subcrop pattern of the Westphalian against the Base Permian Unconformity already clearly shows the shapes of the much younger West Netherlands Basin, Lower Saxony Basin and Lauwerszee Trough as areas of less uplift and erosion. What is now known as the Texel-IJsselmeer High was part of a larger E-W running uplifted zone. The major Hantum Fault Zone at the western margin of the Lauwerszee Trough was active at this time, juxtaposing Westphalian B and C against Westphalian A. The youngest pre-Permian deposits (Westphalian D to Stephanian) are preserved in the Lower Saxony and West Netherlands basins and in a small erosional remnant on the Cleaver Bank High. These are the areas that experienced least uplift during the Saalian event. Strong uplift occurred to the north of the Dutch sector on the Mid North Sea High and Ringkøbing-Fyn High.

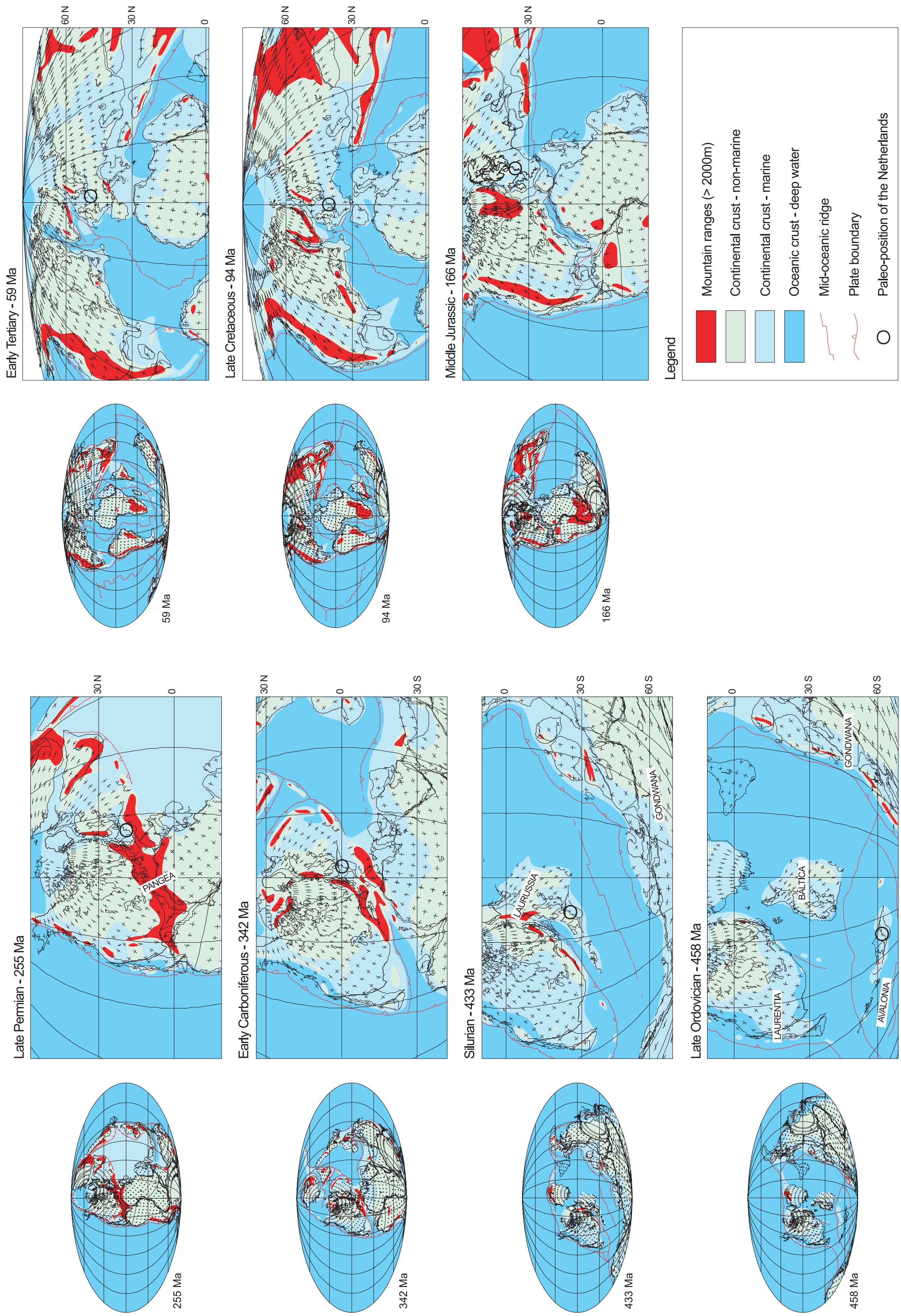
There is surprisingly little evidence of Late Carboniferous or Early Permian rift basins. In northern Germany, narrow fault-bounded basins in which Lower Rotliegend volcanics and clastics accumulated as syn-rift deposits are well documented (Ziegler, 1990a). In the Dutch subsurface, however, the existence of substantial intervals of coeval strata remains speculative. Known occurrences in the Netherlands are restricted to thin volcanic sequences along and within the Ems Low in the easternmost onshore area, and in the Dutch Central Graben and Outer Rough Basin in the northern offshore (Geluk, this volume). High vitrinite reflectances measured in the Westphalian in some wells on the Groningen High (Kettel, 1983), in combination with data from apatite fission-track analysis, suggest that several local heat pulses affected this area. It seems likely that at least one of these occurred prior to the main Early Permian erosion, and caused the high reflectances now occurring directly below the Base Permian Unconformity.

Regional subsidence, induced by the decay of the thermal anomaly that was introduced during the Permo-Carboniferous tectonic-magmatic event, commenced during the Permian, when Rotliegend clastics began to be deposited in the large E-W trending Southern Permian Basin (Van Wees et al., 2000). By then, the area of the Netherlands was located within the arid climate zone at ca. 20° north (Fig. 4). The Southern Permian Basin was bordered to the south by the Variscan Mountains and the stable London-Brabant Massif, and to the north by the Mid North Sea and Ringkøbing-Fyn highs. Clastic sediments entered the basin from all sides, but the main influx was from the south, where the Variscan mountain belt was progressively degraded. The Rotliegend clastics comprise conglomerates and eolian and fluviatile sandstones, passing northward and stratigraphically upward into lacustrine silts and claystones, with in the north also intercalations of evaporites. The forerunners of the Dutch Central Graben, Broad Fourteens Basin and Lauwerszee Trough are already subtly expressed in Rotliegend isopach patterns. Local syn-depositional normal faulting caused minor thickness variations of up to a few tens of metres across faults in the Upper Rotliegend Group.

With subsidence rates exceeding sedimentation rates, a landlocked depression, located well below global sea levels, developed under an arid climate. Early during the Late Permian this depression was catastrophically flooded by saline sea-waters, that upon cyclic evaporation left the halite-dominated Zechstein sequence that attains more than 1500 m in thickness in the axial parts of the Southern Permian Basin (Ziegler, 1988, 1990a).

Mesozoic, Kimmerian breaking-up of Pangea

Rifting related to the Mesozoic break-up of the Pangea supercontinent commenced during the Triassic in the Arctic-North Atlantic and between Greenland and Scandinavia, and slowly propagated southwards into the Central Atlantic domain along the line of future continen-



Late Permian - 255 Ma
255 Ma
PANGEA
30 N
0
Early Carboniferous - 342 Ma
342 Ma
30 N
0
30 S
Silurian - 433 Ma
433 Ma
LAURUSSIA
GONDWANA
0
30 S
60 S
Late Ordovician - 458 Ma
458 Ma
LAURENTIA
BALTICA
AVALONIA
GONDWANA
0
30 S
60 S
Early Tertiary - 59 Ma
59 Ma
60 N
30 N
0
Late Cretaceous - 94 Ma
94 Ma
60 N
30 N
0
Middle Jurassic - 166 Ma
166 Ma
30 N
0
Legend
Mountain ranges (> 2000m)
Continental crust - non-marine
Continental crust - marine
Oceanic crust - deep water
Mid-oceanic ridge
Plate boundary
Paleo-position of the Netherlands

Fig. 4. Global plate reconstructions from the Ordovician to the Tertiary (courtesy Shell). During the Paleozoic the supercontinent of Pangea formed through the collisions of Laurentia, Baltica and Gondwana, and the Caledonian and Variscan orogens formed along the plate sutures. The breaking up of Pangea during the Mesozoic resulted in the formation of rift basins. The Alpine collision of Africa and Europe, caused inversion of these basins and compressional reactivation of pre-existing extensional faults. Since the Early Paleozoic, the area of what is now the Netherlands drifted from the southern hemisphere to its present location, crossing several different climate zones.

tal break-up. An eastern branch of crustal extension propagated during the Early Triassic into the North Sea area (Ziegler, 1988, 1990a). During the Middle Triassic, the extension had reached the southern North Sea, although the degree of extension rapidly decreased southwards. Continued extension in the western rift branch resulted during the Middle Jurassic in continental break-up and the opening of the Central Atlantic Ocean. Towards the end of the Early Cretaceous, crustal separation was achieved in the North Atlantic, and rifting began to concentrate on the area of the Norwegian and Greenland seas (Ziegler, 1988, 1990a). By this time, rifting in the Netherlands effectively ceased. While Europe and North America broke apart, rifting in the Mediterranean domain culminated in crustal separation between Europe and Africa, thus forming a part of the Tethys Ocean.

During the Triassic to Early Cretaceous continental break-up, the area of the Netherlands moved from the arid climate zone to sub-tropical latitudes of the northern hemisphere (Fig. 4).

The Triassic and Early Jurassic sedimentation in the Netherlands took place under continuing thermal subsidence. This led to broad, regular facies patterns, interrupted locally by salt movement which first became widespread after the Early Triassic. Fine-grained clastics of the Lower Triassic Lower Buntsandstein were deposited in brackish to saline lacustrine environments, and were followed by the coarser-grained fluviatile and eolian clastics of the Main Buntsandstein derived from the south and splitting up northward into several thinner sandstone units. During the Middle Triassic, marine environments took over, leading to deposition of the claystones, carbonates and evaporites of the Röt and Muschelkalk formations. Triassic tectonic events are well documented north of the Mid North Sea High. The Horn Graben, which cuts the Ringkøbing-Fyn High and extends into the Southern Permian Basin in the German offshore, is one of the few significant Triassic grabens south of the Mid North Sea High. In the Dutch subsurface, Triassic sequences thicken into the areas of the later Dutch Central Graben and Broad Fourteens Basin. In the West Netherlands Basin, to the south of the Zechstein evaporite basin, the Triassic shows only minor thickness variations across faults. Further north, where the Zechstein salt reaches great thicknesses, it is difficult to link faults in its Triassic cover to sub-salt faults. Thickness variations show that most early salt swells formed during deposition of the Triassic Solling, Röt and Muschelkalk formations, and that piercing salt domes and associated impressive rim-synclines developed later, during the depo-

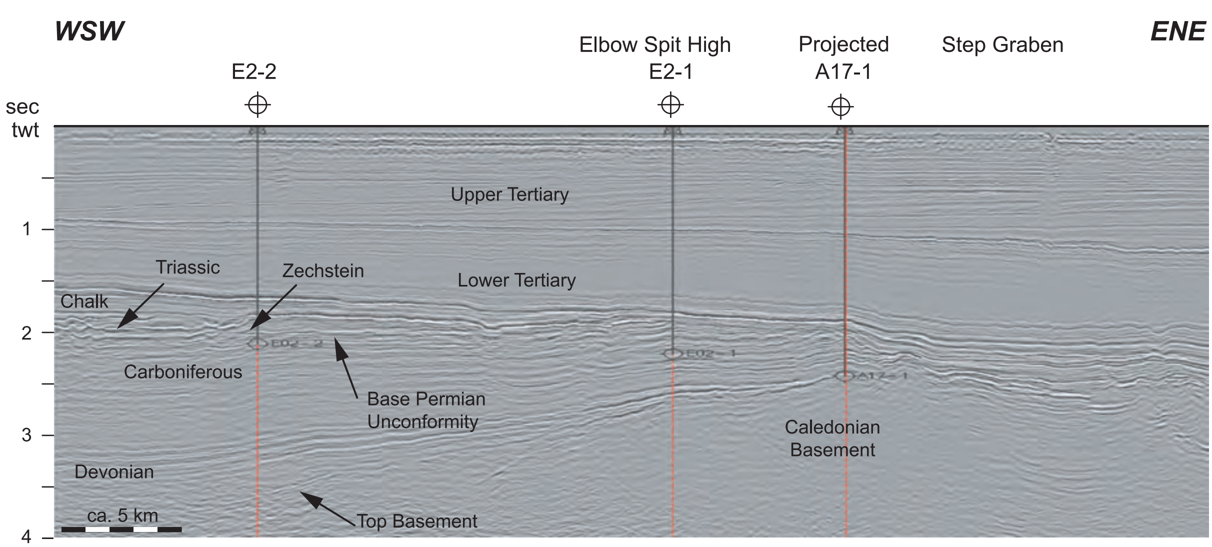


Fig. 5. Seismic line across the Elbow Spit High. Crystalline basement is onlapped by Devonian deposits. For location see Fig. 1.

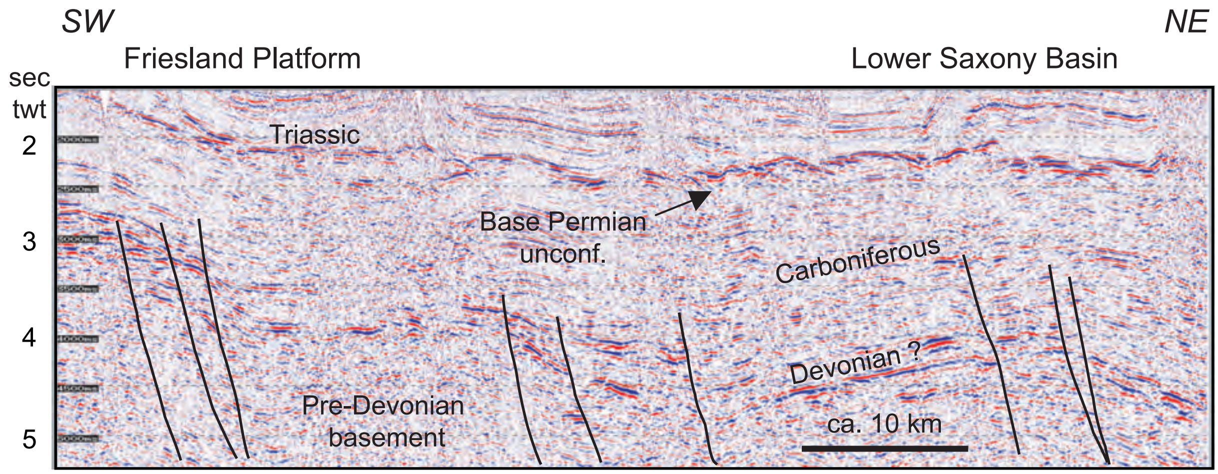


Fig. 6. Deep seismic line across the eastern Netherlands. At pre-Carboniferous levels several major tilted fault blocks indicate significant early extensional faulting. For location see Fig. 1.

sition of the mainly fine-grained coastal-plain to marine clastics of the Keuper Formation. The onset of halokinesis is likely to have been triggered by tectonic activity, in particular where salt 'walls' follow major sub-salt faults (Remmelts, 1995, 1996). Thus, despite the problems of attributing faulting to basement tectonics in areas of thick Zechstein salt, it is likely that Triassic extensional faulting did take place, with decreasing intensity to the south. No heat-flow anomaly nor any igneous activity seems to have accompanied this Triassic extension in the southern North Sea, and no uplift of rift shoulders can be demonstrated.

Truncation of Lower Triassic deposits below the regional Hardegsen Unconformity at the base of the Solling Formation indicates differential uplift during the transition from Early to Middle Triassic in the south of the Southern Permian Basin. The main uplift occurred in the Netherlands Swell in the area of the future Texel-IJsselmeer High. No significant faulting is associated with this phase of uplift.

The Late Triassic and Early Jurassic were periods of relative tectonic quiescence. Some faulting continued in the Dutch Central Graben and locally in the Broad Fourteens Basin, but otherwise slow regional subsidence prevailed. During the Early Jurassic a wide epicontinental sea had developed in which locally more than 1800 m of fine-grained clastics of the Altena Group accumulated, with at the end of the Early Jurassic, deposition of the organic-rich shales of the Posidonia Formation.

During the Middle Jurassic, much of the Dutch offshore area was uplifted in conjunction with the development of the thermal Central North Sea Dome (Ziegler, 1990a; Underhill & Partington, 1993), with sedimentation being restricted to the rift basins. During the Callovian and Oxfordian this dome began to subside again. Following the Middle Jurassic crustal separation in the Central Atlantic and Tethys domains, rifting accelerated in the North Sea rift system, comprising the Viking and Central grabens of Norway and the UK, the Danish Tail End Graben and the Dutch Central Graben (Fig. 4). Southwards the N-S trend terminated against the pervasive NW-SE structural trend that was already established during mid-Paleozoic times, and crustal-extension strain was distributed over a broad area, causing transtensional development of the NW-SE trending Broad Fourteens, West Netherlands, Central Netherlands and Vlieland basins and the Roer Valley Graben. Also the UK Sole Pit Basin and the largely German Lower Saxony Basin resulted from this extension. Thus, the main tectonic elements in the subsurface of the Netherlands were formed mainly during Late Jurassic and Early Cretaceous, Late Kimmerian rifting (Fig. 1). These extensional basins were filled with locally more than 2500 m of sands and clays which, depending on the basin involved, belong to the largely continental Schieland Group, the mainly marine Scruff Group and the continental to restricted marine Niedersachsen Group. Simultaneously, the adjacent platforms were uplifted and deeply truncated. This culminated in the widely recognised Late Kimmerian Unconformity of Early Cretaceous age, which originated during a regional low-stand of sea level. Following this a large open-marine basin formed with deposition of the Rijnland Group: up to 1400 m of mainly fine-grained clastics with coastal sandstones. Comparison of the burial diagrams of the Dutch Central Graben as an example of a Kimmerian rift basin, and of the Cleaver Bank High as an example of a platform area, highlights the difference in burial history (Fig. 7).

Although it can be assumed that regional extension in the Netherlands was in general E-W directed, only the Dutch Central Graben is aligned to conform to this di-

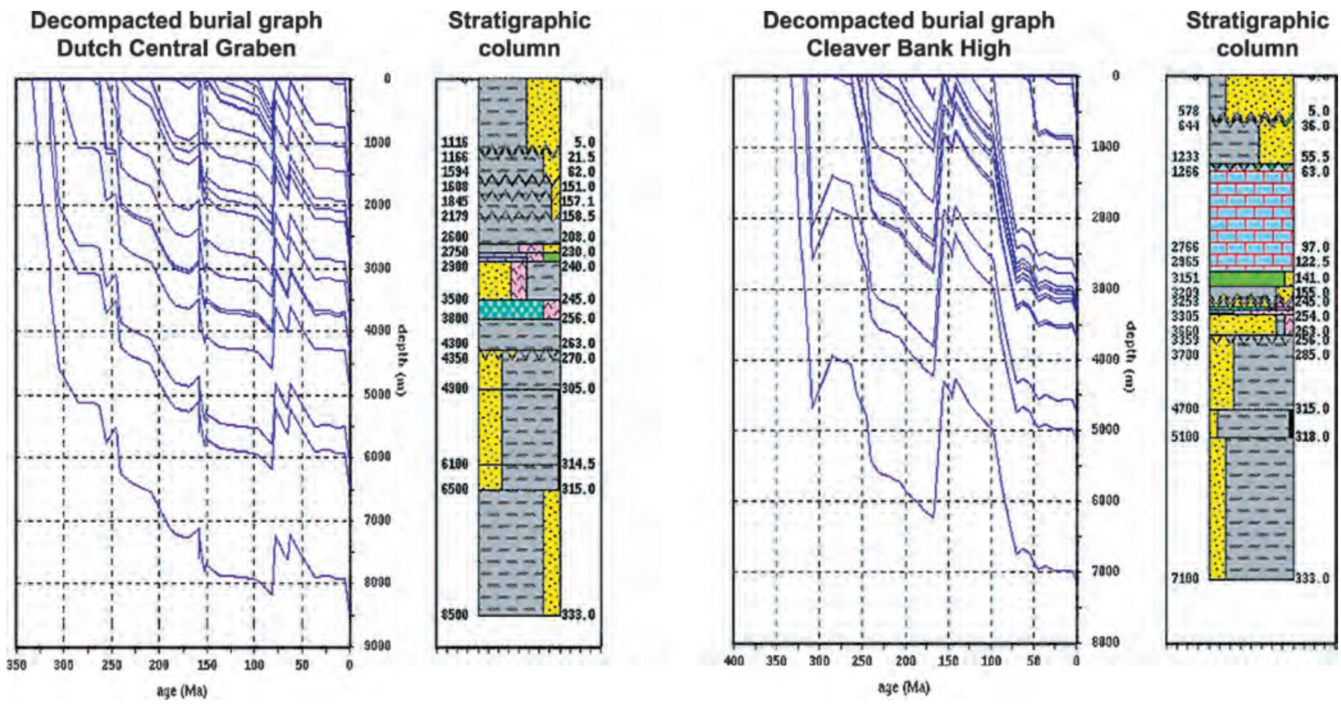


Fig. 7. Burial graphs. (a) Dutch Central Graben, showing accelerated subsidence during the Late Jurassic to Early Cretaceous Kimmerian rifting (160 to 80 Ma) and two pulses of basin inversion (uplift) during the Late Cretaceous and Early Tertiary (80 to 60 Ma). Note strong subsidence during the Quaternary. (b) Cleaver Bank High, showing uplift of this platform area during the rifting from 160 to 80 Ma ago, and no indications of Late Cretaceous uplift.

rection. The NW-SE trending transtensional basins follow older structural trends, which implies that most of their main faults are dextrally reactivated older faults with transtensional displacement. As the main fault systems were repeatedly reactivated under different stress regimes, fault patterns do generally not show unambiguous evidence of oblique movements. Only locally does the fault pattern seem to clearly indicate dextral displacements, such as the Rifgronden Fault Zone between the Terschelling Basin and Schill Grund High.

Late Cretaceous to Early Tertiary, Alpine inversion

Following the mid-Cretaceous opening of the North Atlantic, extensional stresses were progressively concentrated on the area between Greenland, the British Isles and Norway, and tectonic activity abated rapidly in the area of the Netherlands (Ziegler, 1988, 1990a). The Late Cretaceous was characterised by regional thermal subsidence and rising sea levels, with younger sediments overstepping the depositional limits of older sequences, until the entire area of the Netherlands was again submerged in a shallow sea. The up to 1500 m of chalk deposited in this sea forms the only significant carbonate unit in the area since the Triassic.

During the Late Cretaceous, Africa-Arabia began to converge with Eurasia and the Tethys system of oceanic basins started to close. This resulted in the gradual development of the Alpine orogenic system. During the Late Cretaceous and Paleocene, increasing stresses were exerted by the evolving Alpine orogen onto its northern foreland, inducing the inversion of Mesozoic extensional basins and apparently impeding the crustal separation between Greenland and Norway, which was effectuated only at the end of the Paleocene when these compressional stresses abated (Ziegler, 1988, 1990a). Inversion-related uplift of the basins resulted in depositional thinning and in erosion of the Upper Cretaceous chalk and Lower Tertiary clastics, as well as in local truncation of older sediments. On the flanks of the basins, thinning of Upper Cretaceous or Lower Tertiary deposits towards the inversion axes shows uplift to have been a continuous process, albeit with several acceleration pulses. These pulses seem to have been simultaneous in most inverted basins (Fig. 8; De Jager, 2003).

1. Inversion started at the onset of the Late Cretaceous and peaked during the Campanian Subhercynian pulse. Chalk sequences thin towards the inversion axes, with post-Campanian chalk onlapping unconformities in the Broad Fourteens and Central and West Netherlands basins, the Dutch Central Graben and the Roer Valley Graben (Gras, 1995; Gras & Geluk, 1999).
2. In most basins the strongest inversion took place dur-

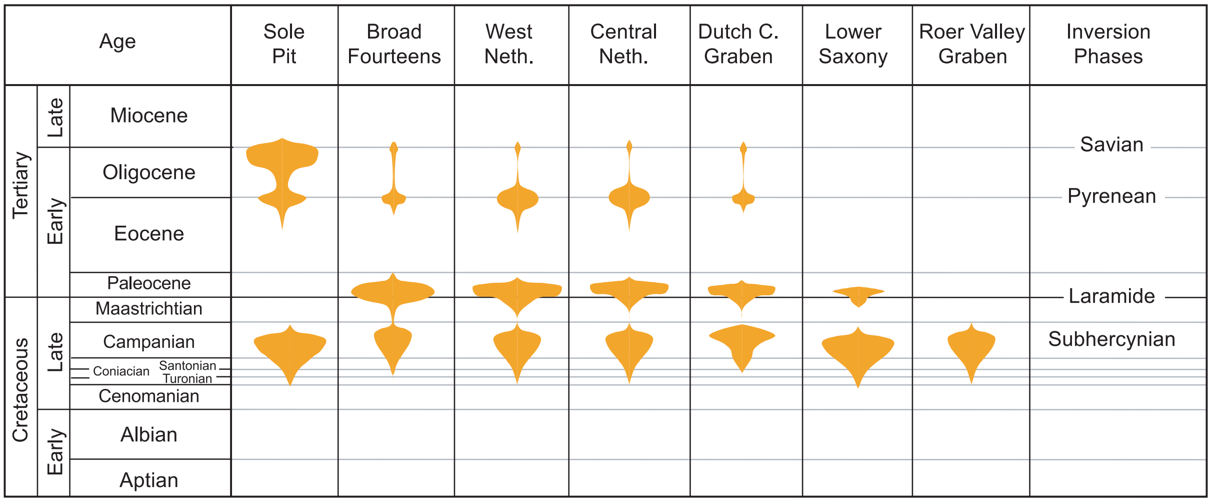


Fig. 8. Timing and relative intensity of Alpine inversion in the Dutch basins and Sole Pit Basin. Note that inversion pulses occurred simultaneously in different basins, but with different magnitudes. Only the Subhercynian pulse appears to have affected all basins.

ing the Paleocene Laramide pulse; it marked the end of chalk deposition and the beginning of deposition of the siliciclastic North Sea Supergroup. Significant erosion down to Jurassic deposits occurred in the West Netherlands, Central Netherlands and Broad Fourteens basins, as well as in the Dutch Central Graben.

3. At the end of the Eocene the Pyrenean pulse caused broad uplift of the West and Central Netherlands basins. The amplitude of this uplift decreases sharply towards the north-west in the Broad Fourteens Basin.
4. An inversion pulse at the end of the Oligocene caused considerable uplift in the UK Sole Pit Basin (Van Hoorn, 1987), but no significant inversion can be demonstrated at that time in the Netherlands.

The deformation related to these pulses varied in intensity from basin to basin, and not all pulses can be recognised in all inverted basins (Figs 8, 9). In some cases an older inversion unconformity appears to have been removed by erosion related to a later phase of uplift. In other cases there is little doubt that an inversion pulse has left a basin unaffected. The Laramide and Pyrenean inversion phases, for example, did not or hardly affect the Lower Saxony Basin (Baldschuhn et al., 1991).

The presence of Zechstein salt had a marked influence on how basin inversion manifested itself. In the West Netherlands Basin, where no Zechstein salt is present, pre-existing faults were reversely reactivated forming prominent ridges of flower structures (Fig. 10). In the Dutch Central Graben, with more than 1 km of Zechstein salt, faults above and below the salt are entirely detached, resulting in a broad uplift of post-salt deposits (Fig. 11). Along the north-east flank of the Broad Fourteens Basin, where the Zechstein salt is at most 200 m thick, the salt acted as a detachment surface along which the basin-fill was thrust onto the adjacent platform (Fig. 12; Hooper et al., 1995; Nalpas et al., 1995).

Detailed analyses of vitrinite-reflectance, fission-track and fluid-inclusion data indicate that the total uplift during the inversion nowhere exceeded 2 km, and in most places amounted to only 1 to 1.5 km (Fig. 7a). This is in disagreement with a reported inversion of up to 3.5 km for some Dutch basins (Dronkers & Mrozek, 1991). It seems that these higher estimates do not fully account for syn-inversion depositional thinning of Upper Cretaceous to Lower Tertiary sequences over the inverted basins (De Jager, 2003).

Late Tertiary to recent structural events

During the Tertiary, rifting in the Lower Rhine Graben propagated northwards into the Netherlands, affecting areas up to the Dutch coast (Ziegler, 1994). The Roer Valley Graben was downfaulted, and up to 2000 m of Tertiary clastics accumulated (Van Adrichem Boogaert & Kouwe, 1993–1997; Van den Berg, 1994; Van den Berg et al., 1994). Recent earthquakes along the main bounding faults indicate that tectonic activity continues to the present day (Dost & Haak, this volume). In the southern North Sea Basin, westward prograding deltaic sediments supplied by the Eridanos River, which drained the Baltic, Fennoscandian Shield area, were deposited during the Neogene and Pleistocene.

Quaternary deposits to the north of the Dutch offshore are up to 1000 m thick, thinning towards the south. For the Dutch offshore and north-western onshore, burial curves show a sharp increase in the rate of subsidence during the last few million years (Figs 7a, b). This increase is considered to be caused by flexural downwarping of the rigid continental crust of the North Sea area in response to the build-up of a NNW-SSE directed compressional stress field (Kooi et al., 1989). While much of the North Sea area still subsides, the south-eastern part of the Netherlands rises in conjunction with uplift of the Rhenish Massif.

An overall review of the Miocene and younger neotectonic activity in the Netherlands was recently published by Van Balen et al. (2005).

Structural elements

Basins in the northern offshore and Waddenzee

The Dutch Central Graben, the southernmost element of the Mesozoic North Sea rift system, is flanked by the shallower Step Graben and Terschelling and Vlieland basins (Figs 1, 2a). The Proto-Dutch Central Graben was probably already present during Late Devonian times, when it formed a sea-way extending from the south into Laurussia (Ziegler, 1990a). There are indications in the Westphalian units subcropping at the Base Permian Unconformity that the N-S trend of the graben was also a structural feature at the end of the Carboniferous. It can furthermore be recognised in the isopach pattern of the Permian Upper Rotliegend Group. Early Permian volcanics are known from its flanks (Geluk, this volume). Major salt walls, which started developing during the Triassic, closely follow the boundary faults of the Mesozoic rift basins in the northern offshore. The development of these salt walls was probably triggered by movements along the faults themselves, suggesting that downfaulting of the Dutch Central Graben was well underway already during Middle Triassic times (Remmelts, 1995, 1996). The increased thicknesses of formations within the graben indicate that differential fault-controlled subsidence continued during Early to Middle Jurassic times (Fig. 11). In particular the Middle Jurassic is thickly developed in the northern sector of the Dutch Central Graben (Heybroek, 1975). The main rifting, however, occurred during the Late Jurassic and Early Cretaceous, affecting northern areas somewhat earlier than areas further south. Thick, fluvial to shallow-marine sequences accumulated within the graben, while adjacent highs and platforms were uplifted and eroded (Fig. 7).

The Step Graben and Terschelling Basin were initially uplifted during the Mid Kimmerian event, leading to erosion of older Jurassic and Triassic deposits. These shallower basins subsided later and to a lesser extent than the Dutch Central Graben, resulting in thinner Upper Jurassic sequences than in the graben. The Terschelling Basin is delineated by major faults. The Hantum Fault Zone along its southern margin was already established during the Early Permian. While there is less convincing evidence that the Rifgronden Fault Zone at the northern margin of the basin is also an old fault, its parallelism to the Hantum Fault Zone and its unusual direction with respect to

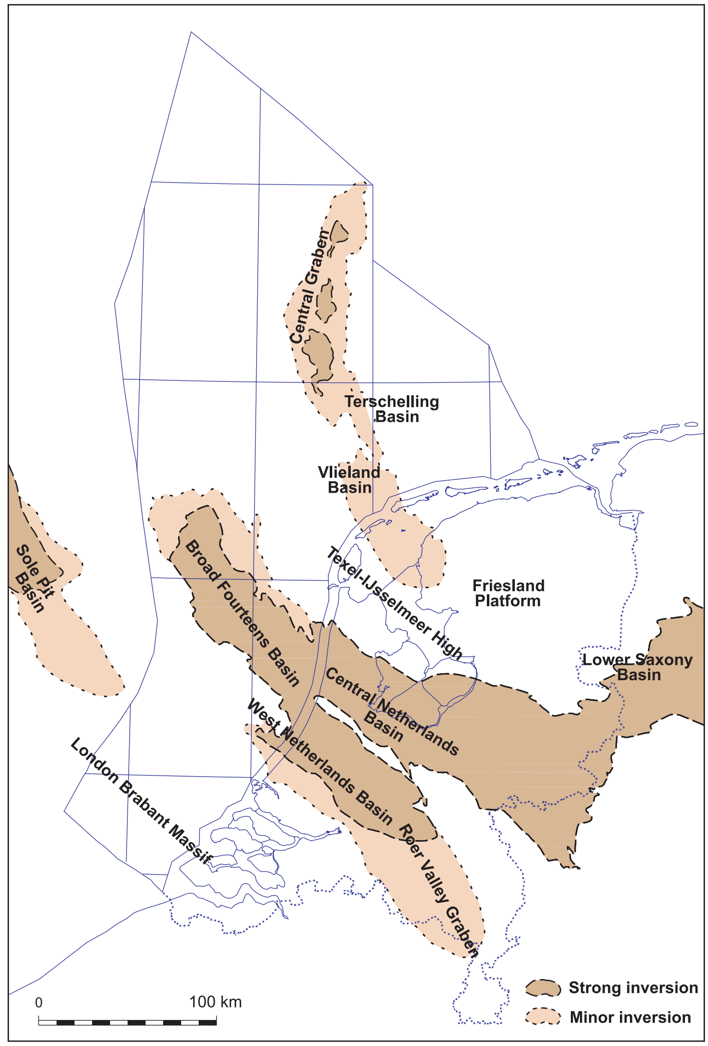


Fig. 9. Inverted basins in the Netherlands. Light shading shows mildly inverted areas, where the Upper Cretaceous chalk is thin as a result of erosion and/or limited deposition; dark shading shows strongly inverted areas, where no chalk has been preserved.

the Kimmerian E-W extension speak in favour of its origin as a reactivated pre-existing fault.

The development of the Vlieland Basin has been related to dextral shear along a NW-SE trend, linking the southern termination of the Dutch Central Graben with the Lower Saxony Basin (Herngreen et al., 1991). A significant igneous event associated with Late Kimmerian rifting led to the Zuidwal Volcano, a major alkaline volcanic complex that developed in the Vlieland Basin. This complex is expressed by a marked magnetic anomaly, and almost 1 km of volcanic rocks have been penetrated by well Zuidwal-1, which appears to have drilled into the volcanic neck (Perrot & Van der Poel, 1987). The local temperature anomaly that must have been associated with this igneous activity, seems to have limited the Late Jurassic subsidence, and only thin Upper Jurassic

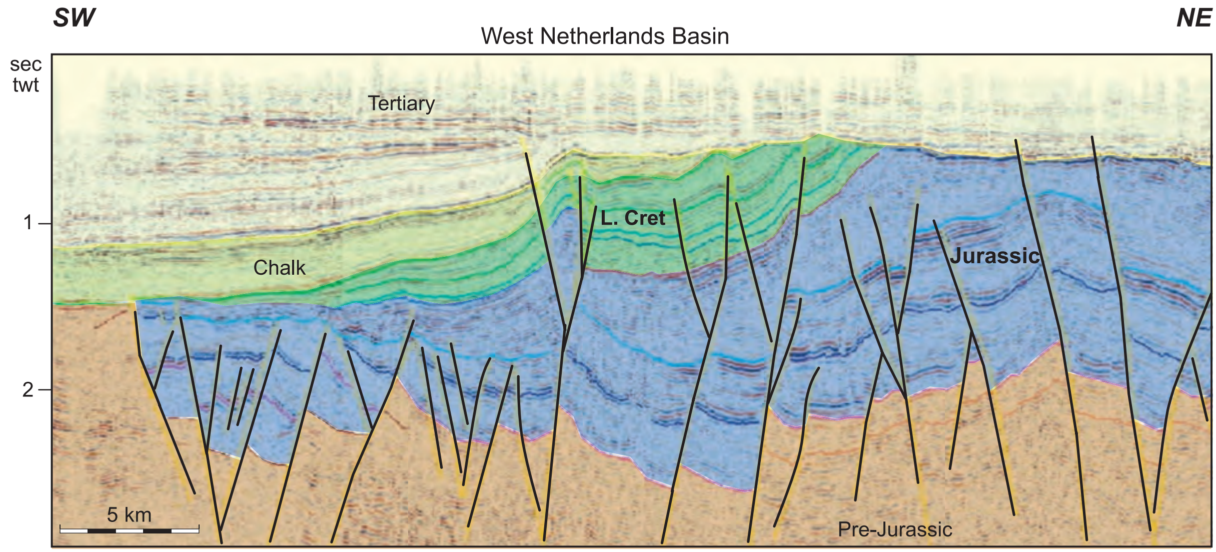


Fig. 10. Seismic section running from the south-western flank of the West Netherlands Basin into the inverted basin. For location see Fig. 1. Note the increased thickness of the Jurassic in the basin, and how the Lower Cretaceous onlaps the south-western basin margin. Chalk deposits show depositional thinning towards the inversion axis, and are truncated by the Base Lower North Sea Group Unconformity. During the Late Cretaceous inversion, flower structures originated through transpressional reactivation of pre-existing extensional rift faults. Oligocene inversion caused broad basin uplift, without significant faulting.

deposits are overlain by Lower Cretaceous post-rift sediments.

Inversion affected the Dutch Central Graben with decreasing intensity towards the north. The Subhercynian pulse resulted in significant erosion of pre-Campanian deposits. Within the most strongly inverted areas this erosion probably already cut down to the Lower Cretaceous Vlieland formations. Upper Cretaceous chalk overlies the eroded surface. A locally developed intra-chalk sandstone at the level of the Campanian unconformity (well F17-4) may represent the erosional products of Lower Cretaceous, and possibly even Jurassic clastics. Seismic sections show that the main inversion occurred during the Paleocene Laramide pulse. Older deposits are deeply truncated below the unconformably overlying Paleogene Lower North Sea Group. During the Early Tertiary, the Dutch Central Graben subsided less than its surroundings, resulting in a much thinner Lower Tertiary sequence; no marked unconformities can be observed in this sequence on good-quality seismic data.

Basins in the south and west of the Netherlands

The Broad Fourteens, West Netherlands and Central Netherlands basins and the Roer Valley Graben form a

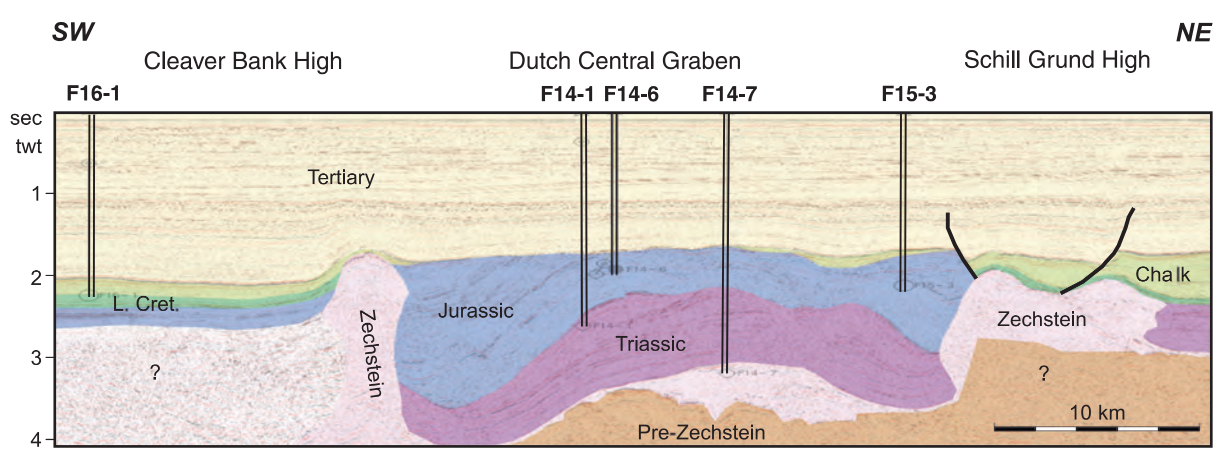


Fig. 11. Seismic section across the inverted Dutch Central Graben. For location see Fig. 1.

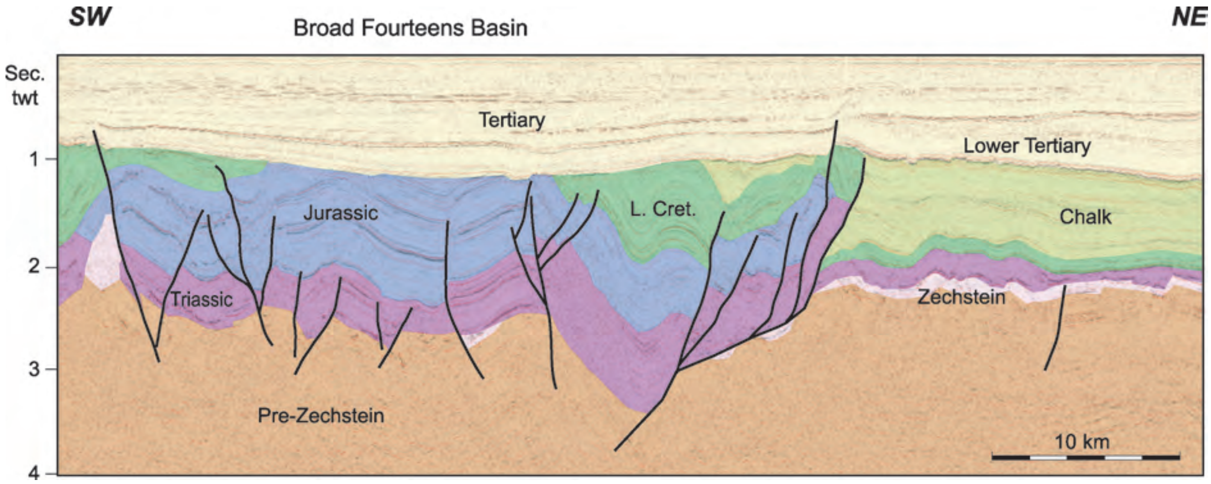


Fig. 12. Seismic section across the northern sector of the Broad Fourteens Basin. For location see Fig. 1. Note that the Jurassic is thick within the inverted basin, and absent on the adjacent platform. The inversion resulted in an impressive thrust fault along the north-east flank of the basin. Contrary to the West Netherlands Basin, the Oligocene inversion pulse did not affect this part of the Broad Fourteens Basin.

NW-SE trending contiguous set of Mesozoic extensional basins (Fig. 1). The boundary between the West and Central Netherlands basins is formed by the NW-SE striking Mid Netherlands Fault Zone, which comprises the Peel Block, Maasbommel High and Zandvoort Ridge, and extends into the offshore IJmuiden High, and ultimately into the Indefatigable Fault Zone in British waters. As a result of significant Cenozoic inversion and erosion, the structural evolution of this fault zone is not easily reconstructed. It is already expressed in the subcrop pattern of Westphalian units below the Base Permian Unconformity, and there is no doubt that it is a major structural feature that has been repeatedly reactivated during much of the geological history. There is no clear boundary between the Central Netherlands and Broad Fourteens basins, nor between the West Netherlands Basin and Roer Valley Graben, which trend in slightly different directions. The NW-SE orientation of the Mesozoic extensional basins in the south and west of the Netherlands does not conform to the assumed E-W direction of extension during Late Kimmerian rifting. With the exception of the Broad Fourteens Basin, these basins are sub-parallel to the subcrop pattern of the Westphalian at the Base Permian Unconformity. Thus their localisation was probably controlled by pre-existing pervasive structural elements. Similar to the Dutch Central Graben, increased thicknesses of Triassic sediments in the Broad Fourteens Basin and Roer Valley Graben indicate early establishment of these basins. Seismic sections show evidence for local, Early to Middle Jurassic normal faulting in the Broad Fourteens and West Netherlands basins.

In the West Netherlands Basin, which is devoid of Zechstein salt, Late Kimmerian extension resulted in a series of half-grabens filled with Upper Jurassic and Lower Cretaceous syn-rift clastics of the Delfland Subgroup. Younger Cretaceous post-rift deposits onlap the basin margins (Figs 2c, 10; Bodenhausen & Ott, 1981; De Jager et al., 1996). Igneous activity accompanied the rifting, as is evidenced by volcanic rocks and intrusive sills near the base of the Delfland Subgroup. During the Late Cretaceous and Early Tertiary inversion, pre-existing faults were reactivated reversely, with transpressional movements forming a series of prominent NW-SE trending flower structures. Along many individual faults, a reversed sense of movement can still clearly be deduced: normal offsets at deeper, and reverse offsets at shallower stratigraphic levels. Thinning of the Upper Cretaceous chalk from the south-east towards the inverted basin, with several internal unconformities visible on seismic data, clearly shows that the flower structures were formed during the Late Cretaceous Subhercynian inversion. During the Paleocene Laramide inversion, the West and Central Netherlands basins were both further uplifted and deeply eroded, locally down to the Triassic. The Subhercynian inversion in the Central Netherlands Basin is less well documented, as due to deep erosion during later phases no Upper Cretaceous deposits are preserved, but also there in several wells Danian chalk overlies pre-Cretaceous successions (NITG, 2004b). During the Early Tertiary, inversion of this basin continued. At the south-west margin of the West Netherlands Basin, the Lower Tertiary thins towards the inverted basin (Fig. 10). During the Oligocene, erosion took place in the most inverted sector of the basin, as well as in the Central Netherlands Basin. Noteworthy is that, although the Subhercynian and Laramide inversions resulted in reverse reactivation of pre-existing faults, the Pyrenean inversion caused broad basin uplift without fault reactivation. At the end of the inversion movement, relaxation of the stress regime resulted locally in normal reactivation of faults.

The area of the Roer Valley Graben was also affected by the Subhercynian inversion. The effects of this inversion were strongest in the north-west, while the adjacent Maasbommel High subsided considerably (Geluk et al., 1994; Gras, 1995; Gras & Geluk, 1999).

The Broad Fourteens Basin was affected by the same structural events as the Central and West Netherlands basins, but displays a different structural style (Fig. 2b; Van Wijhe, 1987a, b). The main difference in style is caused by the presence of Zechstein salt in the northern half of the basin. Below this salt, at the level of the Rotliegend, the structuration resulted in a series of tilted fault blocks (Fig. 12). As structuration above the salt was decoupled from sub-salt faulting, inversion of the half-grabens filled with Upper Jurassic sediments, characteristic of the West Netherlands Basin, is less clearly developed. The Laramide inversion resulted in the development of broad NW-SE trending anticlinal arches (Dronkers & Mrozek, 1991; Nalpas et al., 1995). The Eocene Pyrenean inversion affected mainly the southern sector of the basin.

Due to the late, Eocene to Oligocene, inversion of the West and Central Netherlands basins, their Tertiary cover is thin, locally not more than 500 m. This contrasts sharply with the Tertiary of the northern sector of the Broad Fourteens Basin, which is twice as thick. In the Roer Valley Graben, the Tertiary reaches 2000 m (Geluk et al., 1994; Van den Berg, 1994; Van den Berg et al., 1994). Subsidence was particularly pronounced after the Oligocene, when crustal extension associated with the development of the Lower Rhine rift system propagated into the Netherlands. The Peel Boundary Fault, separating the graben from the Peel Block to the north-east, shows a post-inversion throw of up to 1 km.

Basins in the north-east of the Netherlands

The Lower Saxony Basin is an E-W trending, Late Jurassic and Early Cretaceous, transtensional basin, largely located within Germany; its westernmost part extends just into the Netherlands, with Jurassic and Cretaceous deposits thinning westwards onto the Friesland Platform. To the north the basin is practically connected with the Lauwerszee Trough, which is located immediately west of the Groningen High. Both the Dutch sector of the Lower Saxony Basin and the Lauwerszee Trough lack half-grabens and tilted fault blocks. The outlines of both clearly conform to subcrop patterns below the Base Permian Unconformity, indicating that their development was controlled by pre-existing structural trends. This is also clear from the significant Hantum and Gronau fault zones, located west of the Lauwerszee Trough and south-west of the Lower Saxony Basin respectively, which offset Westphalian units at the Base Permian Unconformity. The Gronau Fault Zone continued to be active in the Cenozoic (Van Adrichem Boogaert & Kouwe, 1993–1997).

At Base Rotliegend level, the Lauwerszee Trough is only several hundreds of metres deeper than the adjacent Friesland Platform and Groningen High, and is not a very significant basin. It is, however, a distinct and long-lived feature, not only expressed in subcrop patterns, but also in increased thicknesses of Rotliegend, Lower Cretaceous and Paleogene deposits. The latter may in part also be due to withdrawal of Zechstein salt into flanking salt walls, which have grown significantly during Early Cretaceous and Tertiary times.

The Ems Low along the German-Dutch border flanks the Groningen High to the east. Middle Permian igneous rocks within the graben represent a western extension of much more widespread igneous activity in Germany, although they are somewhat younger (Geluk, this volume). This activity was associated with crustal extension that led to a fan-shaped system of Early Permian grabens filled with Lower Rotliegend volcanogenic clastics in the area of Hannover (Ziegler, 1990a). Also the Triassic is thickly developed in the Ems Low, which became overprinted by the Lower Saxony Basin during the Late Kimmerian rifting phase.

Although major Late Cretaceous inversion affected the German sector of the Lower Saxony Basin (Betz et al., 1987; Baldschuhn et al., 1991), its Dutch sector was only inverted mildly.

Highs and platforms

The London-Brabant Massif in the south and the Texel-IJsselmeer High in the centre of the country remained positive structural elements during much of the geological history. The Ringkøbing-Fyn High and Mid North Sea High are located just north of the Dutch offshore area, and will not be discussed here. Platforms, such as the Cleaver Bank High, the Ameland Block and the Schill Grund High, have a more complete sedimentary record than the aforementioned highs. The difference with the basins which they flank is that instead of accumulating thick syn-rift sediments, they were uplifted and eroded during the Middle Jurassic and Early Cretaceous rifting phases (Fig. 7).

After sedimentation encroached upon the London-Brabant Massif during the Westphalian, the massif received sediments only for a relatively short period during Late Cretaceous to mid-Tertiary times (Ziegler, 1990a). Any deposits that might have accumulated along its flanks during other periods were eroded.

The Texel-IJsselmeer High became strongly uplifted during the Early Permian, Saalian tectonic event (Rijkers & Geluk, 1996), as is clearly indicated by the Westphalian subcrop pattern below the Base Permian Unconformity. No rocks of Late Permian to Middle Jurassic age have been preserved on this high. Thin deposits of this period may have been laid down, but during Late Kimmerian rifting the high was uplifted, and eroded down to the Westphalian. Regional post-rift subsidence affected the high, and from Late Cretaceous times onwards it subsided and sediments were being deposited. As a result of repeated uplift and erosion the Albian Holland Formation now overlies Westphalian A, with a hiatus of no less than 220 Ma (Rijkers & Geluk, 1996).

During long periods the platform areas were sites of sedimentation, similar to the Mesozoic extensional basins. Only during the Mid and Late Kimmerian rifting pulses, changes occurred. The platforms were uplifted and became subject to erosion. Consequently, Upper Jurassic or Lower Cretaceous syn-rift deposits are not present, and Triassic and older series are unconformably overlain by a thin veneer of post-rift Lower Cretaceous and a relatively thick sequence of Upper Cretaceous chalk. Only very little of the Upper Triassic to Middle Jurassic deposits was preserved, and locally erosion cut down to the Zechstein. The removal of these deposits was so effective that it becomes difficult to imagine that they were ever deposited, particularly where no strong angular unconformity occurs between the Cretaceous post-rift and the eroded older sequences. In the Mesozoic rift basins, Upper Triassic to Middle Jurassic series reach thicknesses of hundreds of metres to more than 1 km. On the platforms, these deposits were probably not quite as thick, as early differential subsidence of the rift basins appears to have started already during the Triassic. Nevertheless, hundreds of metres of Triassic and Lower to Middle Jurassic sediments were probably deposited on most platforms (Fig. 7b). This is indicated by the complete absence of near-shore deposits within the Lower Jurassic of the Netherlands, and also by the uniform thickness of, for example, the Toarcian Posidonia Shale. Wherever this unit is preserved, even in Germany, it always is ca. 30 to 35 m thick, and has very similar log expressions and oil-source-rock qualities. Had the platforms been areas of non-deposition, greater variations in thickness and lithology would surely have resulted (Ziegler, 1990a).

Although the platforms were not uplifted during the Alpine inversion, Alpine compression did affect the Cleaver Bank High, just north and north-west of the Broad Fourteens Basin. This is evidenced by the occurrence of several pop-up structures at Rotliegend level (Fig. 13). Even though these features occur in areas of thick Zechstein salt, where it is difficult to assess the timing of faulting, their compressional nature unambiguously associates them with the Late Cretaceous and Early Tertiary inversion pulses, as these are the only post-Rotliegend compressional pulses in the Netherlands. In an area of very thin Zechstein salt, in block L13 just north-east of the Broad Fourteens Basin, a subtle but clearly expressed Riedel fault pattern at Base Tertiary level suggests dextral shear during the Pyrenean inversion phase through reactivation of deeper faults (Oudmayer & De Jager, 1993).

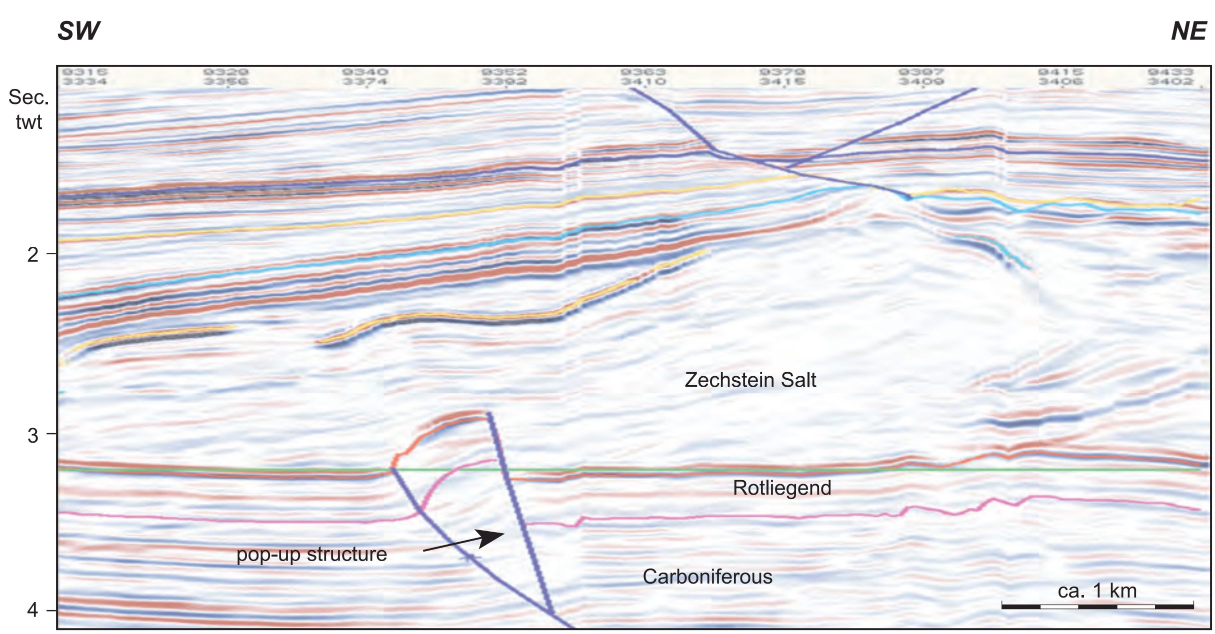


Fig. 13. Seismic section on the Cleaver Bank High, showing at Rotliegend level a NW-SE trending pop-up structure bounded by a reverse fault. The compressional nature of this structure indicates that it must have formed during Late Cretaceous to Early Tertiary inversion. For location see Fig. 14.

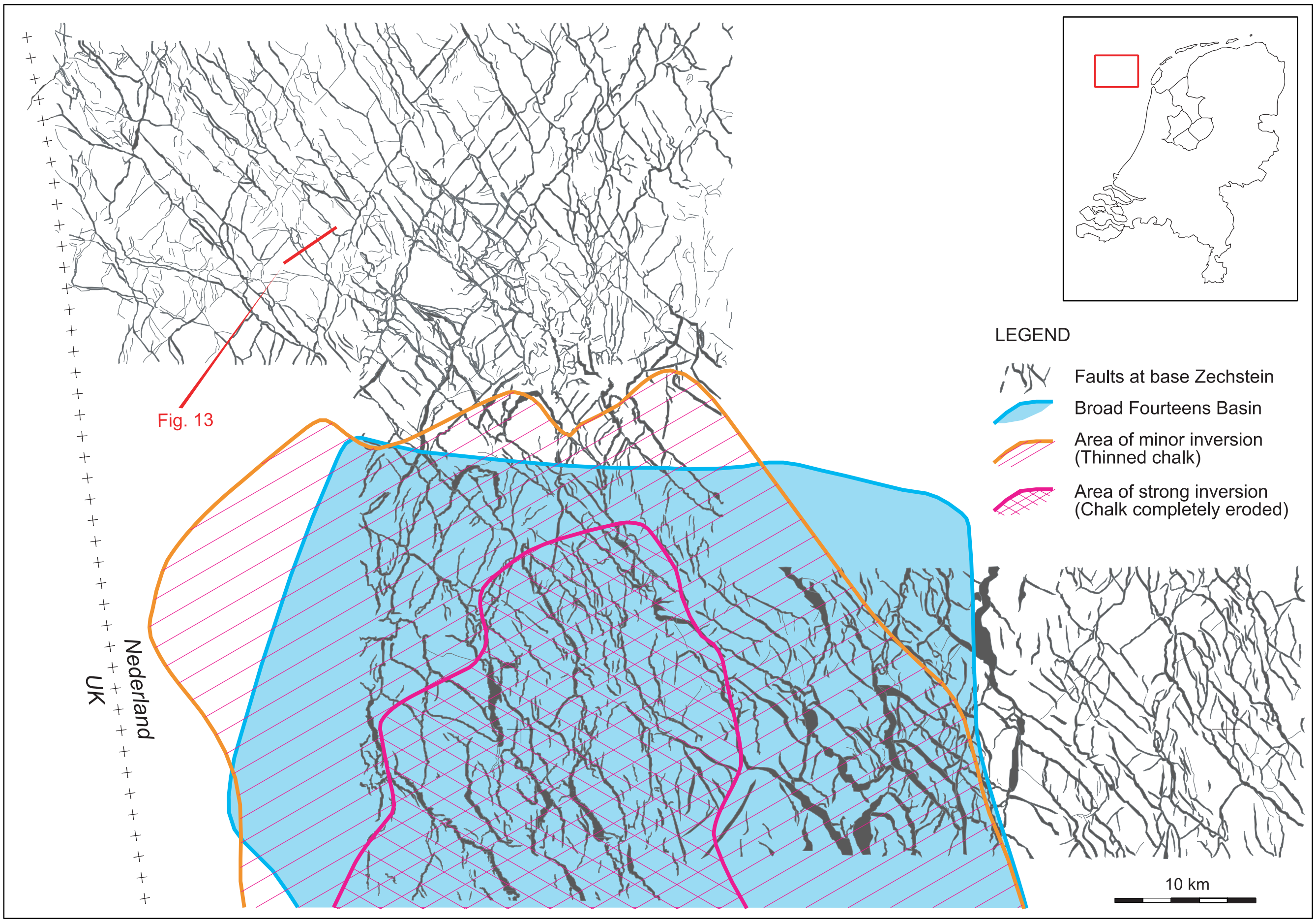


Fig. 14. Fault pattern at Rotliegend level in the northern sector of the Broad Fourteens Basin and adjacent Cleaver Bank High, based on good-quality 3D seismic data. Despite significant differences in structural development, there are no significant differences in fault patterns and fault density between the Broad Fourteens Basin and adjacent platform areas.

Fault patterns

Below the Zechstein salt, at the level of the Rotliegend, many structural elements show a characteristic rhomboid pattern of intersecting fault trends (Figs 14, 15). The dominant family of faults trends NW-SE and probably dates from mid-Paleozoic times when Avalonia, including the London-Brabant Massif, docked against Laurussia. The major Hantum and Mid Netherlands fault zones belong to this family, and show clear evidence of repeated reactivation. The second most common fault trend is aligned NE-SW to ENE-WSW, and became regionally established during the Early Permian (Ziegler, 1988, 1990a). Since both trends were thus already in place prior to Rotliegend and Zechstein deposition, their presence in these younger intervals results from fault reactivation. The fault intersections generally do not indicate consistent offset geometries, and it is likely that both trends were reactivated several times more or less simultaneously. In the northern offshore, the NW-SE fault trend becomes less important and N-S faults paralleling the Dutch Central Graben and Step Graben dominate.

In the northern sector of the Broad Fourteens Basin and on adjacent platform areas, 3D seismic data clearly show NW-SE and NE-SW fault trends at Rotliegend level, with secondary faults trending WNW-ESE and N-S (Fig. 14). Apart from the better imaging of these faults on the platforms, there is no significant difference between the basin and the platforms in the directions and spacing of faults. This is remarkable in view of their different structural histories. Apparently, during rifting and later inversion, pre-existing faults were reactivated, but only few new faults formed at Rotliegend level. Pop-up structures on the Cleaver Bank High, north and north-west of the Broad Fourteens Basin, follow the NW-SE, NE-SW and WNW-ESE trends, and show that those three trends were reactivated simultaneously during Late Cretaceous and Early Tertiary compression (Fig. 13).

In the north-east of the Netherlands, the main fault

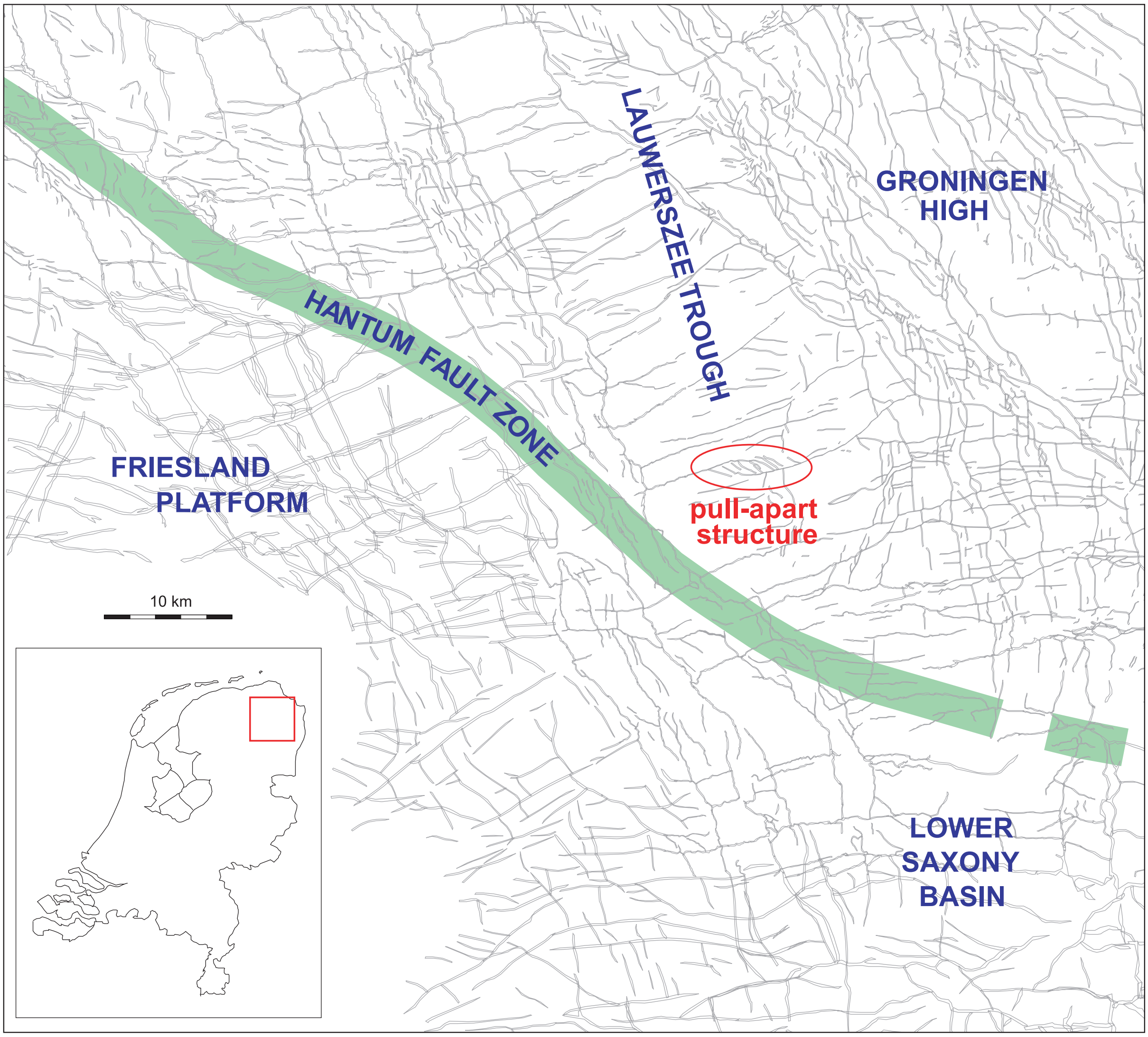


Fig. 15. Fault pattern at Rotliegend level, north-east Netherlands. The major Hantum Fault Zone cannot be mapped as a through-going fault on good-quality 3D seismic data. Note the presence of a small but crisply displayed pull-apart feature within the southern Lauwerszee Trough, indicating local dextral transtensional displacement.

trends in the Zechstein and Rotliegend appear to have rotated slightly as compared to the western offshore: the faults trend mainly NNW-SSE and ENE-WSW. In the Lower Saxony Basin, E-W trending faults are common (Fig. 15). As indicated previously, the large NW-SE trending Hantum Fault Zone at the western boundary of the Lauwerszee Trough and the NW-SE trending Gronau Fault Zone along the south-western boundary of the Lower Saxony Basin represent old fault trends. In some cases a considerable lateral displacement along individual faults seems likely, but the amount of strike-slip is mostly not more than that of the vertical offsets: mainly in the order of tens to hundreds of metres for the Lauwerszee Trough area. Fault maps at Cretaceous and Tertiary levels of this area show that both the NNW-SSE and the ENE-WSW trending faults were reactivated during the Tertiary.

Within the West Netherlands Basin, faults affecting the Rotliegend show a characteristic anastomosing pattern of dominant WNW-ESE and NNW-SSE, and secondary N-S and E-W directions (Fig. 16). Dextral displacement along the main faults would be consistent with E-W Late Kimmerian extension as well as with N-S Late Cretaceous and Early Tertiary compression. However, in many of the inverted structures the Riedel patterns at the base of the Cretaceous rather appear to indicate sinistral shear. Also for the Pijnacker oil field in the West Netherlands Basin, an Early Cretaceous sinistral movement along WNW-ESE faults has been postulated (Racero-Baena & Drake, 1996). A dextral displace-

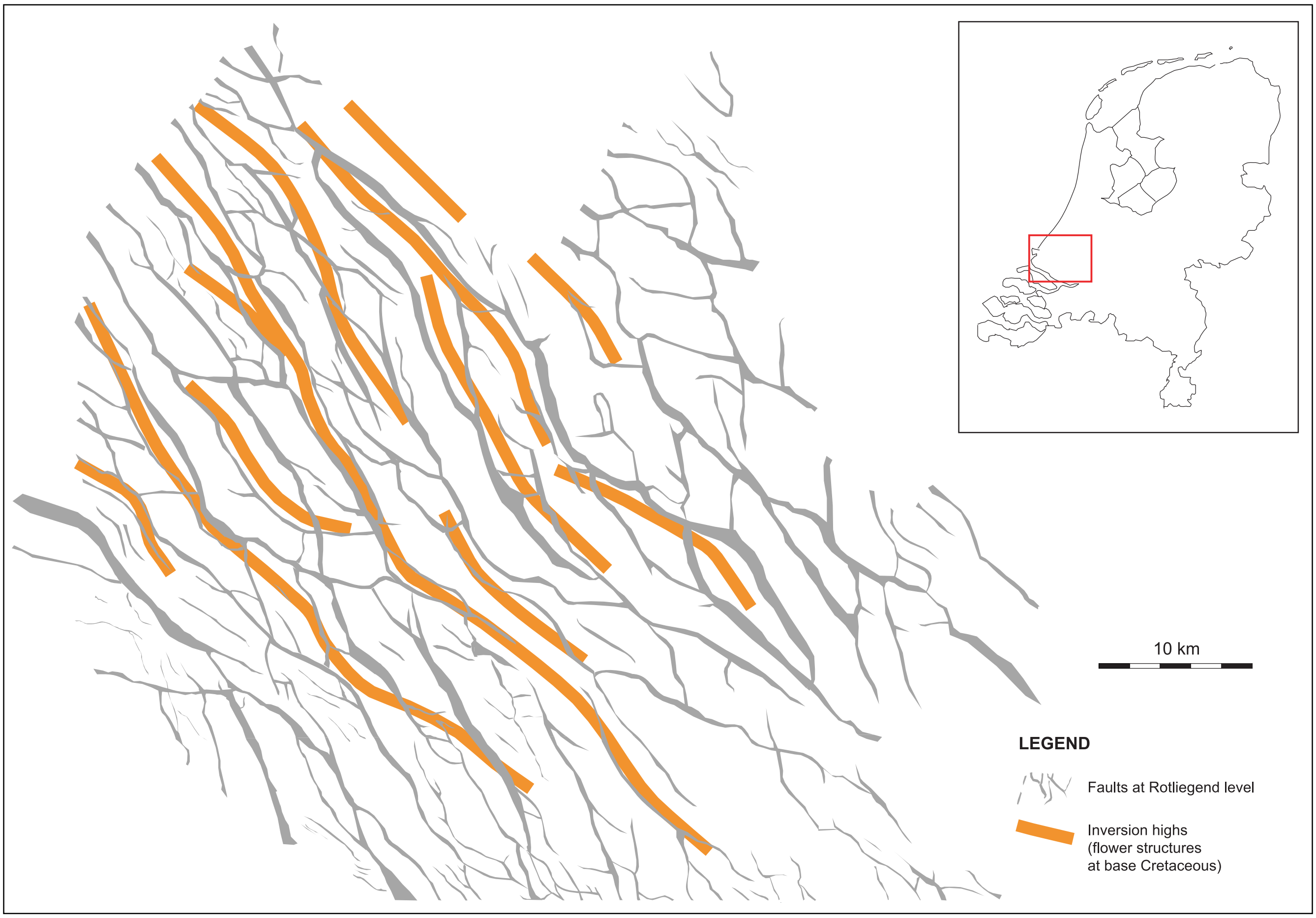


Fig. 16. Fault pattern at Rotliegend level, West Netherlands Basin, based on good-quality 3D seismic data. The rhomboid fault pattern has dominant WNW-ESE and NNW-SSE faults. Continuous flower structures at base Cretaceous, caused by Late Cretaceous to Early Tertiary inversion, zig-zag from one fault trend to the other, and must have formed through simultaneous reactivation of both trends.

ment during subsequent Alpine inversion, interpreted for this field, agrees with the regional model. However, it appears that it may not be appropriate to interpret regional stress directions from these repeatedly reactivated fault geometries. In fact, the Late Cretaceous compressional flower-structures were formed through reactivation of both NNW-SSE and WNW-ESE fault trends. What looks like an alignment of continuous flower structures at Cretaceous levels, can be seen to follow different fault trends at Rotliegend level, zigzagging from one to the other (Fig. 16). Apparently, none of the pre-existing fault trends was ideally oriented to be preferentially reactivated. Therefore present-day fault trends cannot be readily interpreted in terms of paleo-stress directions. As reliable indicators to determine the accumulated amount, or even the sense, of lateral displacement are rare, it is impossible to unravel the sense and amount of displacement during the various evolutionary phases of individual faults.

Overpressures

Formation pressures in the Netherlands vary from hydrostatic to highly overpressured. Overpressures occur in most stratigraphic intervals below the Neogene Upper North Sea Group. They generally increase with depth and towards the north. In the West Netherlands, Central Netherlands and Broad Fourteens basins all reservoir pressures are essentially hydrostatic, with only locally minor overpressures of up to several tens of bars. In the north-east Netherlands, overpressures occur in the Triassic, and northward increasing overpressures occur below the Zechstein salt in the Rotliegend and Carboniferous. In the northern offshore, overpressures have been measured in most stratigraphic intervals, with particularly high values, close to minimum horizontal stress, where the Triassic sandstone units and the Rotliegend are deeply buried. In addition, hard overpressures occur in the 'Zechstein floaters', slabs of platy dolomite, limestone and anhydrite that became involved in salt diapirism, and that are now 'floating' within Zechstein salt, completely detached from other porous rocks.

The Upper Cretaceous and Danian Chalk Group is overpressured in the northern offshore. The overpressures increase to 50 to 70 bars above hydrostatic in the far northern Dutch offshore, and even to 100 to 150 bars in the Danish and Norwegian sectors of the North Sea respectively (Fig. 17a). Overpressures in Lower Cretaceous and Jurassic sequences also increase in the northern offshore, reaching some 80 bars above hydrostatic (Fig. 17b). Regionally, overpressures within the Triassic follow a similar trend as in the younger intervals, reaching 80 bars above hydrostatic in the north. Overpressures of up to 80 or 90 bars are also measured in the Triassic in the Dutch sector of the Lower Saxony Basin, close to the German border (Fig. 17c). A noteworthy occurrence of hard overpressures is evident in Triassic Volpriehausen and Dethfurt sandstones in the area of the Terschelling Basin and southern Dutch Central Graben, where these sandstones occur in effective pressure cells. They are underlain by Zechstein salt, overlain by anhydritic claystones and Triassic evaporites, and flanked by Zechstein salt walls. In these cells, excess pressure cannot leak off, and pressures of locally more than 400 bars above hydrostatic, close to, or at, formation strength, are present. Where the Triassic sandstones are truncated at the Base Cretaceous Unconformity, the overpressures follow the regional Cretaceous-Jurassic pressure trend, in agreement with the notion that only very effective seals can preserve high overpressures. In the Rotliegend, overpressures start developing further south than in the younger intervals. In the north-east, in Groningen and Friesland, Rotliegend overpressures increase northward to about 100 bars above hydrostatic, and initial pressures in the Ameland gas field are even some 170 bars above hydrostatic. Further north, only few pressure data are available, but measured pressures of more than 200 bars above hydrostatic suggest that pressures increase further towards the Terschelling Basin and Dutch Central Graben (Fig. 17d).

World-wide, overpressures often occur in young deltaic settings, where compaction-driven de-watering of shales and sands is impeded by the overall low permeability of the clay-dominated sequence. The setting in the Netherlands is different. Overpressured reservoirs often have complex burial histories, and may today not even be at their maximum burial depths. Modelling the development of overpressures in Triassic pressure cells has shown that compaction and temperature increase during Tertiary burial alone can easily result in the build-up to the observed overpressures. This is in agreement with the observation that overpressures increase towards the north, together with the increasing thickness of the Tertiary overburden. For the preservation of overpressures, however, aquifers need to be effectively and three-dimensionally sealed off. In areas where reservoir communication is possible with surrounding rocks, no or only low overpressures are present.

Summary

The oldest dated rocks in the Netherlands are of Silurian age, and the locally 10 or more kilometres thick stratigraphic record above them is on the whole almost continuous to the present day. The setting of basins, platforms and highs changed repeatedly as a consequence of tectonic processes that accompanied the world-wide interaction of lithospheric plates from the Early Paleozoic onwards. Episodes of compression and mountain building alternated with phases of extension and the creation of rifted basins, while the present-day area of the Netherlands drifted from the southern hemisphere across equatorial regions to its current position in the temperate climate belt around 52° north.

The main plate-tectonic events which influenced the area of the Netherlands were: i) the assembly of the Pangea supercontinent during the Paleozoic through the collisions of Laurentia, Baltica and Gondwana, resulting in the Caledonian and Variscan orogens along plate boundaries, ii) the Kimmerian breaking up of Pangea during the Mesozoic, and the accompanying formation of rift basins, and iii) the Alpine collision of Africa and Europe, causing inversion of the earlier formed Mesozoic rift basins and compressional reactivation of pre-existing extensional faults.

Periods of igneous activity, both intrusive and extrusive, accompanied some of these events. Crustal extension and thinning, accompanied by increasing heat-flow, caused uplift and erosion of platform areas adjacent to rift basins. During post-rift periods of declining heat-flow, regional subsidence occurred, while depositional areas expanded onto previously emerged platform areas.

Despite the many tectonic events, a high degree of fault parallelism can be observed. The general structural model is, therefore, one of repeated reactivation of basement faults which continue to control the structural grain, despite changes in tectonic regime and stress direction. Only rarely and by coincidence did pre-existing structural features have orientations that conformed to the regional stress of later episodes. Consequently, most faults have a component of oblique slip.

In much of the Dutch subsurface, diapirism of the thick Permian Zechstein salt led to impressive salt domes of sometimes more than 3 km height, reaching close to the present-day surface. This thick salt package has caused faulting above the salt to be decoupled from basement faulting, hampering unambiguous assessment of the timing of movements along individual faults at sub-salt levels. However, as many of the salt walls and domes are follow-

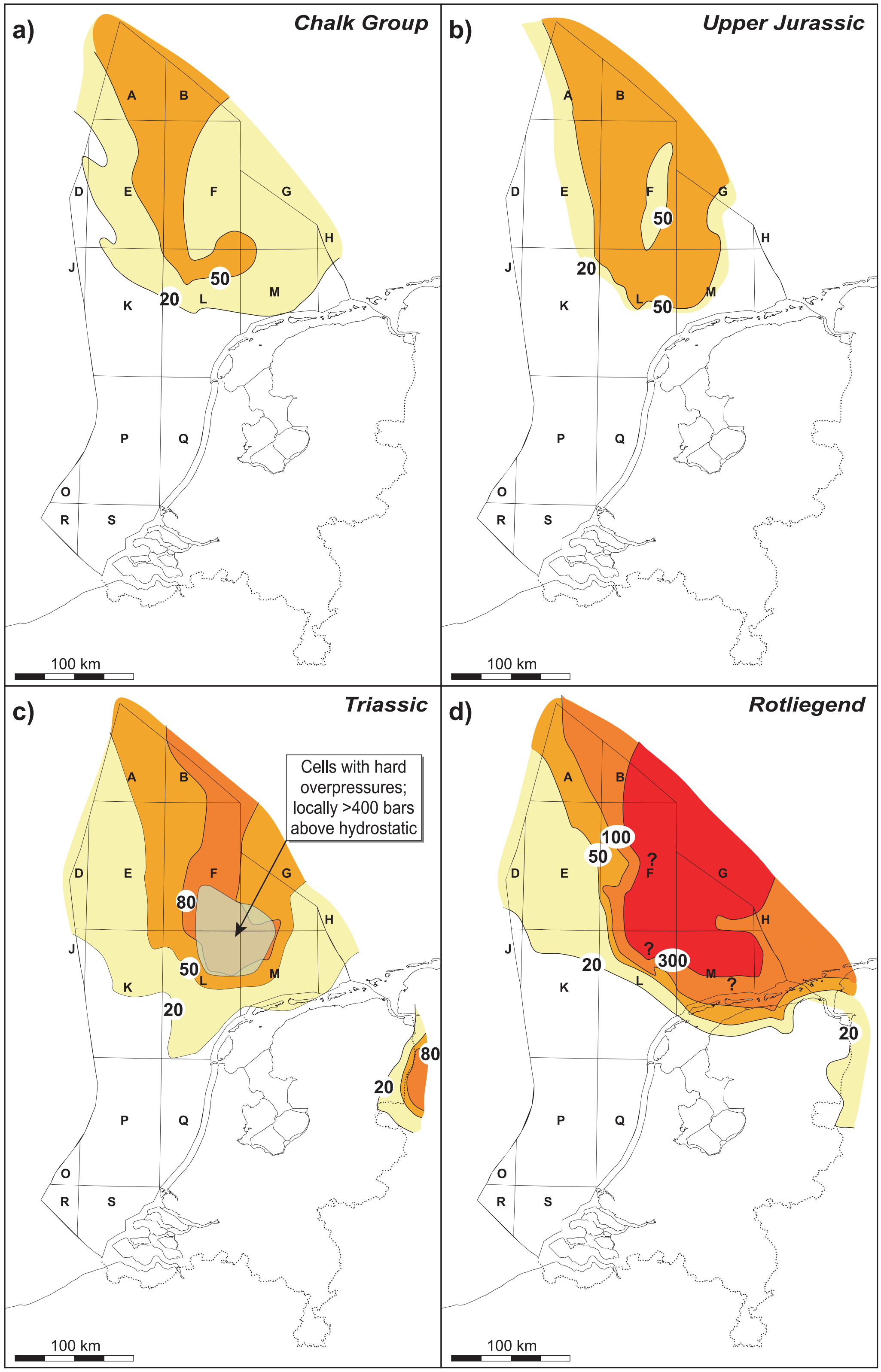


Fig. 17. Overpressures in bars in the Dutch subsurface. (a) Upper Cretaceous Chalk Group. (b) Upper Jurassic. (c) Triassic. In the area of the southern sector of the Dutch Central Graben and the Terschelling Basin, salt walls define cells with overpressures to more than 400 bars above hydrostatic. (d) Middle and Upper Permian Rotliegend.

ing main sub-salt fault trends, the salt movements were apparently triggered by basement faulting.

The tectonic forces which affected the subsurface not only shaped its present-day structural configuration, but also exerted a profound influence on the depositional history.

The overpressures in the subsurface are caused primarily by burial during the Tertiary. Their preservation is strongly related to the presence of effective seals.

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Pre-Silesian

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ABSTRACT

Pre-Silesian rocks in the Netherlands comprise two different units: i) a deeply buried, moderately deformed, slightly metamorphic and scarcely known Caledonian basement of Precambrian to Silurian age, and ii) Middle Devonian to Early Carboniferous siliciclastics and carbonates which unconformably cover this basement. The basement is best known from the Anglo-Brabant deformation belt. Its rocks, marine, shelf to deep-water clastics, were deformed during the three-plate convergence of the Caledonian orogeny (Late Silurian to Early Devonian). The Netherlands was situated on the Gondwana-derived Avalonia plate, which collided with Baltica and Laurentia. Following the last, Early Devonian, phase of the Caledonian orogeny, a horst-and-graben topography controlled the deposition, with fluvial deposits residing on the footwall blocks, and basinal deposits in the hanging-wall blocks. Horst blocks shielded a large portion of the area from siliciclastic influx coming from the Mid North Sea High, allowing a widespread carbonate platform to form during the Early Carboniferous in the central and southern Netherlands. The synsedimentary horst-and-graben faults were gradually overstepped, and gave way to the regional subsidence of a Silesian coal-bearing molasse basin in the Variscan foreland.

Keywords: Netherlands, Lower Paleozoic, Devonian, Lower Carboniferous, tectonics, paleogeography, Caledonian, Acadian

Introduction

Pre-Silesian rocks are known only from a few wells, situated in the southern and northernmost parts of the Netherlands. These wells, in conjunction with data from Belgium, Germany and the United Kingdom, allow hypothetical models to be put forward regarding the pre-Devonian, Caledonian development of the country. For the Devonian and Dinantian more data are available, but models remain hypothetical in view of the large area not covered by data (Fig. 1). The models are based upon integrated well and seismic data and literature; key references include Legrand (1968), Bless et al. (1976, 1983), Ziegler (1989, 1990), De Vos et al. (1993), Pharaoh et al. (1995), Van Grootel et al. (1997), Maynard et al. (1997), Kockel (1998), Lokhorst (1998), Pharaoh (1999) and Verniers et al. (2002).

The top of the pre-Silesian displays a considerable variation in depth. In the south-easternmost tip of the Netherlands, it lies just above the Dutch Ordnance Level (NAP), whereas in the nearby Roer Valley Graben and in the central and northern offshore, it is situated below 9000 m (Fig. 1).

No hydrocarbon accumulations have been encountered so far in the pre-Silesian, but the post-Caledonian part of this succession is still considered as potentially prospective for hydrocarbon exploration (Cameron & Ziegler, 1997; De Jager & Geluk, this volume). In the south-east of the Netherlands, mineral water is produced from Dinantian deposits in well Thermae-2000 for a spa (Fig. 1). In this region, several other wells also encountered mineral water (NITG, 1999). In northern Belgium, the saline aquifer in Dinantian carbonates is used for the extraction of geothermal energy (Beerse-Merksplas doublet) and the Dinantian itself for the storage of natural gas (Heibaart dome in Loenhout). In central and eastern Belgium, various pre-Silesian units are being quarried, of which the dense, bluish-grey Dinantian limestones are best-known.

Stratigraphy and regional correlations

The pre-Silesian succession can be subdivided into (Fig. 2):

- a Precambrian to Silurian, weakly metamorphic, moderately deformed Caledonian succession;
- an overlying post-Caledonian, Middle to Upper Devonian and Lower Carboniferous, mildly deformed succession of sediments.

The Caledonian succession is not defined in the lithostratigraphy of the Netherlands, since it has been encountered in two wells only. In Belgium, numerous wells reach this succession in the Brabant Massif. Essentially, it consists of fine siliciclastics, except for the Lower Cambrian which is coarser; turbidites are common in the Silurian. For detailed descriptions, reference is made to Legrand (1968), Verniers & Van Grootel (1991), De Vos et al. (1993), Servais et al. (1993) and Verniers et al. (2001).

The pre-Silesian post-Caledonian sediments are subdivided into the Devonian Old Red, Kyle and Banjaard groups, and the Lower Carboniferous Farne and Carboniferous Limestone groups (Van Adrichem Boogaert & Kouwe, 1994). The groups within the Devonian, like those within the Lower Carboniferous, are to a large extent lateral equivalents. They are conformably overlain by Silesian, i.e. Upper Carboniferous, siliciclastic sediments.

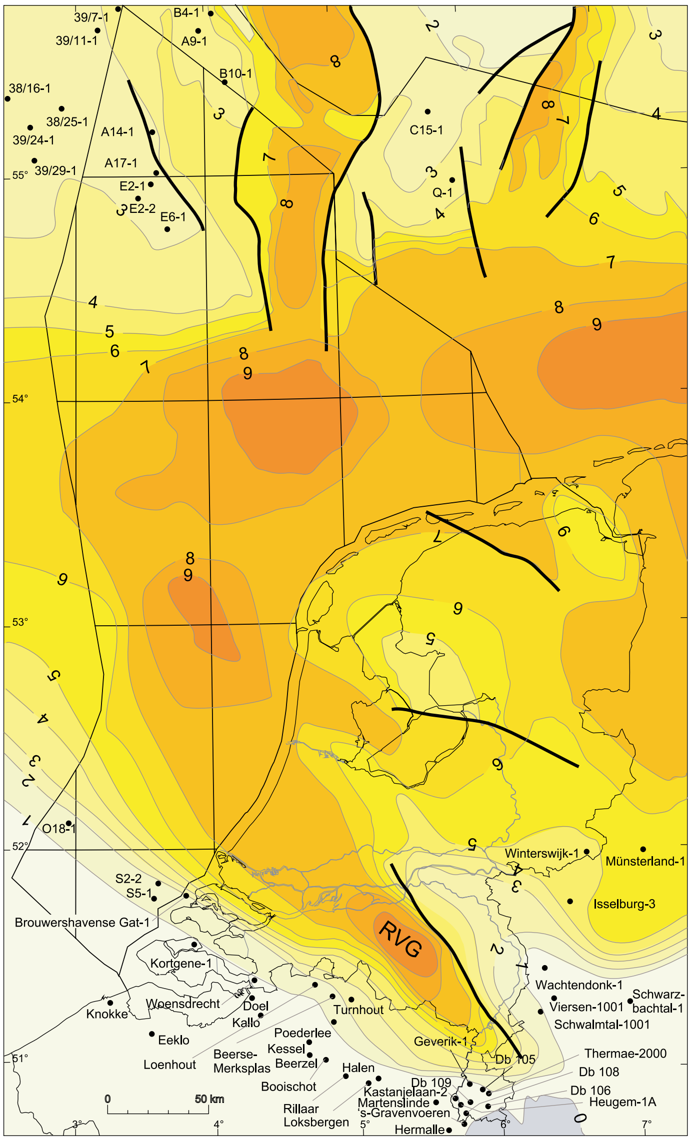

Fig. 1. Schematic depth map in kilometres of the top of the pre-Silesian. For the southern Netherlands and adjacent parts of Belgium and Germany, maps of Bless et al. (1976), Legrand (1968) and NITG (2001) have been used. In other areas, the top of the pre-Silesian has been constructed by adding regional isopachs of the Silesian (Van Buggenum & Den Hartog Jager, this volume) to the depth of the base Permian surface (Lokhorst, 1998). The map further gives an overview of wells which encountered the pre-Silesian. RVG: Roer Valley Graben.

The subdivision of the Devonian and Lower Carboniferous depends heavily on well data and literature from the surrounding countries. Formation and group names have partly been adopted from these countries, e.g. the Kyle, Old Red and Farne groups in the area of the Mid North Sea High (Cameron, 1993), and the Carboniferous Limestone Group in the southern Netherlands (Van Adrichem Boogaert & Kouwe, 1994).

Tectonic development

In pre-Silesian times a large part of the plate-tectonic amalgamation of north-west Europe was completed. The Caledonian orogeny, recently redefined by McKerrow et al. (2000a), represents the closure of the Iapetus and Tornquist oceans between three microcontinents: Laurentia, Baltica and Avalonia. The Netherlands was situated on Avalonia (Figs 3, 4).

On faunal evidence, Cocks & Fortey (1982) showed that the Avalonian microplate was attached during the Cambrian to Gondwana in a southern high-latitude position, and migrated during the Ordovician to a lower latitude near Baltica, to which it shows a similar fauna in the Upper Ordovician. Paleomagnetic data indicate that Avalonia occupied positions at 60° S in Early Ordovician Tremadoc, 40° S in Late Ordovician Caradoc, 20° S in Silurian and 10° S in Late Devonian Famennian times (Torsvik et al., 1996; Cocks, 2000).

In the late Early Ordovician Arenig, the Avalonian crust rifted away from Gondwana, rapidly drifted northwards and joined Baltica in a soft collision or 'docking' event at the end of the Ordovician, closing the Tornquist Ocean. This resulted in uplift of several parts of Avalonia, from the Ardennes to the south of Baltica. The Lower Paleozoic rocks, known from inliers in the Ardennes, were folded during the Ardennian deformation phase between the Middle Ordovician Llanvirn and the Late Silurian (Mansy et al., 1999; Verniers et al., 2002). This phase is contemporaneous with the Shelveian phase in England (McKerrow et al., 2000a). The weakly deformed North German-Polish Caledonides are also related to the collision between Avalonia and Baltica (Katzung et al., 1993; Giese et al., 1997; Dallmeyer et al., 1999). The paleogeographic framework of this collision is still poorly understood; reference is made to the discussions in Pharaoh (1999) and Verniers et al. (2002). Meanwhile the Brabant Through, an elongated sedimentary basin, developed in the southern Netherlands and Belgium, where Silurian deep-water sediments accumulated (Bless et al., 1980; Ziegler, 1990; Verniers & Van Grootel, 1991).

During the Ordovician and Silurian, the oceanic Iapetus plate subducted below Laurentia and Avalonia. In north Wales and in the Lake District in northern England, calc-alkaline magmatism occurred during the Middle to Late Ordovician, testifying to this subduction. The Late Ordovician calc-alkaline magmatic rocks in Anglia and the inferred granites in the Brabant Massif are probably also related to subduction, although the paleogeographic context is less clear in these cases. Deep seismic profiles in the southern North Sea (Blundell et al., 1991) revealed a SW-dipping deep reflector in the mantle below the Dowsing-South Hewett Fault Zone, which reflector was interpreted as a possible remnant of this Ordovician subduction zone (Lee et al., 1993; Pharaoh et al., 1995). The Dowsing-

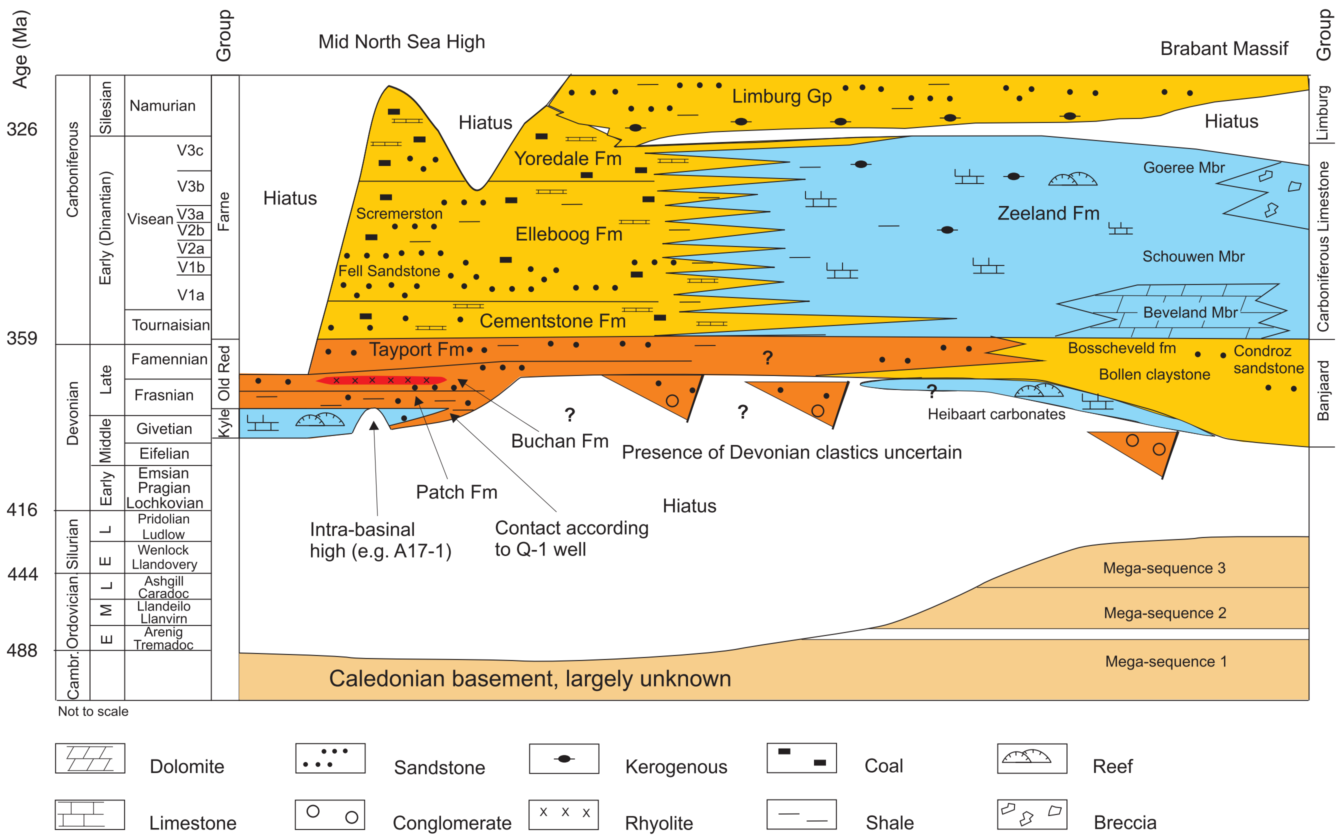

Fig. 2. Schematic stratigraphic overview of the pre-Silesian, Cambrian to Lower Carboniferous deposits in the Netherlands and surrounding areas (after Cameron, 1993; Van Adrichem Boogaert & Kouwe, 1994). Well data used for calibration are only available for the Mid North Sea High and the Brabant Massif. The stratigraphy in the area in between is highly speculative.

South Hewett Fault Zone is possibly the suture between two different terranes within the amalgamated Avalonia microplate (South North Sea-Lüneburg Terrane; Pharaoh, 1999).

The Avalonia-Baltica continent collided with Laurentia in the latest Silurian and Early Devonian. This collision finally closed the Iapetus Ocean and gave rise to a large continent called Laurussia. In Belgium, during this last phase of the Caledonian orogeny, called Acadian phase by Van Grootel et al. (1997) and Brabantian phase by Verniers et al. (2002), the sedimentary fill of the Silurian Brabant Trough was inverted, i.e. folded, faulted and uplifted. To the north-west, this deformed Brabant Massif is connected to the eastern England Caledonides (Pharaoh, 1999). This orogenic belt is referred to as the Anglo-Brabant Massif (Pharaoh et al., 1993), the Anglo-Brabant fold belt (Van Grootel et al., 1997) or the Anglo-Brabant deformation belt (Verniers et al., 2002; Fig. 3).

The name London-Brabant Massif, used in the literature on the post-Caledonian, refers to the area including both the Proterozoic Midlands Microcraton with its cover of undeformed Paleozoic, and the folded Anglo-Brabant deformation belt. It represents the stable area against which the Variscan orogeny abutted.

The Acadian deformation phase affected not only the Anglo-Brabant deformation belt but also the English Lake District, Wales, southern Ireland and the Canadian Appalachians (McKerrow, 1988). All these areas show an angular unconformity between the Upper Silurian and the Middle Devonian. On the Brabant Massif, little deformed Middle Devonian strata overlie stronger deformed Lower Paleozoic metasediments. The structure of this massif shows a WNW-ESE oriented Cambrian core curving eastwards to an E-W direction and surrounded on both sides by Ordovician and Silurian rocks (De Vos et al., 1993).

A low-grade epizonal metamorphism affects most of the massif, except its south-western part. There is no systematic difference in metamorphic grade between Cambrian, Ordovician and Silurian rocks, but towards the north and the south this grade diminishes (Van Grootel et al., 1997), regardless of the ages of the rocks involved; this decrease is sharper in the south.

As outcrops appear only in some river valleys in the south of the Brabant Massif, subcrop mapping was car-

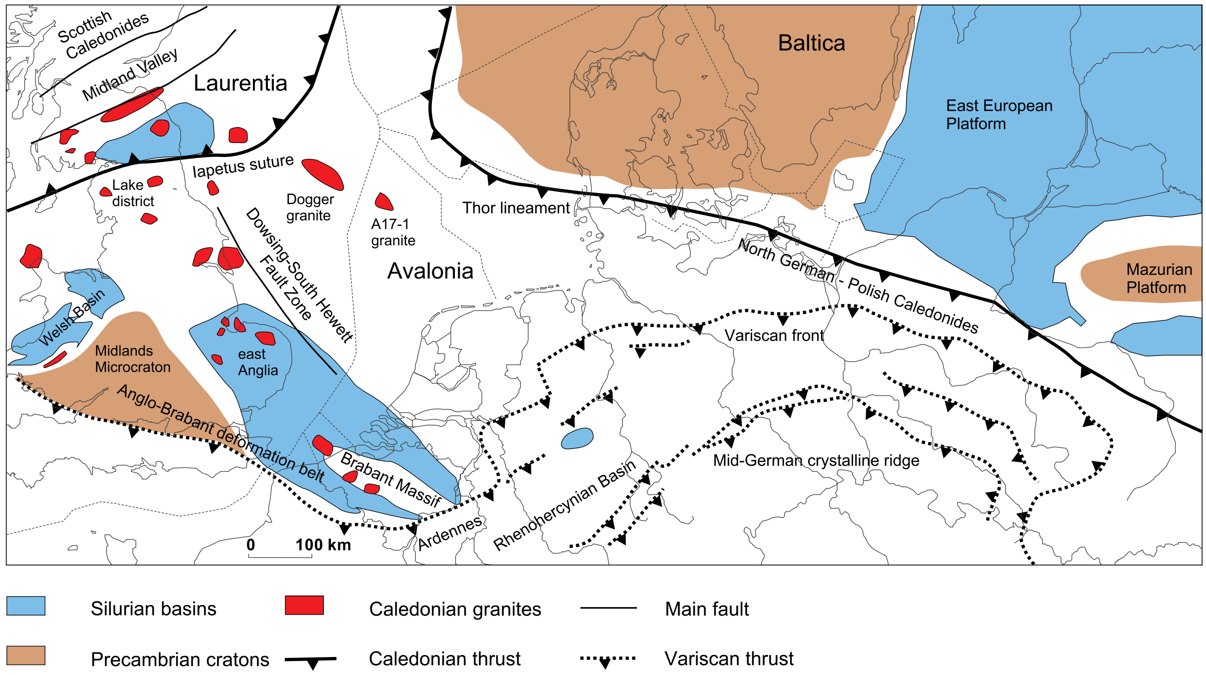

Fig. 3. Structural overview of north-west Europe showing the paleo-continents and the main Caledonian and Variscan structures (after Ziegler, 1990; Pharaoh et al., 1995; Kockel, 1998).

ried out based on extensive borehole data, helped by the interpretation of aeromagnetic and gravity maps (De Vos et al., 1993; Everaerts et al., 1996; Mansy et al., 1999). The Lower Cambrian Tubize formation, consisting of decimetre to metre-thick turbiditic fining-upwards layers of greenish chlorite-rich sandstone, siltstone and phyllite, contains disseminated magnetite in many layers. Its magnetic properties reveal numerous structures on large and intermediate scales (Sintubin, 1997, 1999).

A series of negative gravity anomalies are mostly interpreted as granitic bodies (Everaerts et al., 1996; Mansy et al., 1999), or simply referred to as low-density bodies at depth. These bodies had a significant influence on the Acadian structural development of the Brabant Massif (De Vos, 1997; Sintubin, 1999; Debacker, 2001; Verniers et al., 2002).

During the northward migration of Avalonia in the Ordovician, the Rheic Ocean opened between Avalonia and Gondwana (Ziegler, 1990; Fig. 4). After the collision with Baltica, i.e. during the latest Ordovician and earliest Devonian, subduction of the Rheic oceanic plate occurred, both southwards under the Armorican terranes and northwards under Avalonia (Franke, 2000). The inferred granites of the Brabant Massif could be related to early northward subduction of this plate.

In the Devonian and Carboniferous, the three-plate convergence gave way to a two-plate convergence (Ziegler, 1989, 1990; McKerrow et al., 2000b). The latter involved the northwards drift of Armorican terranes and Gondwana against Laurussia, with a major sinistral translation between Laurentia-Greenland and Fennoscandia-Baltica. South of the Avalonian area the Rheic Ocean closed, resulting during the Late Carboniferous in the nappe complex of the Variscan Mountains (Ziegler, 1990; Franke, 2000). The thrust front was situated south of the Brabant Massif in the Midi-Aachen Thrust. The Rhenohercynian fold-and-thrust belt, which includes the Ardennes and the Rhenish Massif, has an Avalonian crust and Avalonian faunas in the Lower Paleozoic, and is considered to have been thrust upon more northerly cratonic areas of Avalonia during the Late Carboniferous (Pharaoh, 1999; Franke, 2000; Oncken et al., 2000).

The Devonian and Early Carboniferous basin development in the Netherlands occurred in response to back-arc extension in the Rhenohercynian Basin to the south-east of the Netherlands (Ziegler, 1990). The result was a series of WNW-ESE trending, fault-bounded, half-grabens in the southern North Sea, similar to the basins described in the English onshore (Leeder, 1988; Fraser & Gawthorpe, 1990; Chadwick, 1993, Hollywood & Whorlow, 1993). The basins are separated by granite-cored horsts such as encountered in the well A17-1 (Ziegler, 1990). Sedimentation resumed here during the late Middle Devonian. Strong Middle Devonian to Early Carboniferous block faulting has been documented from both the northern flank of the London-Brabant Massif (Fig. 5; Muchez & Langenaeker, 1993) and the Mid North Sea High (Quirk, 1993; Maynard & Dunay, 1999). Similar block-faulting is thought to ex-

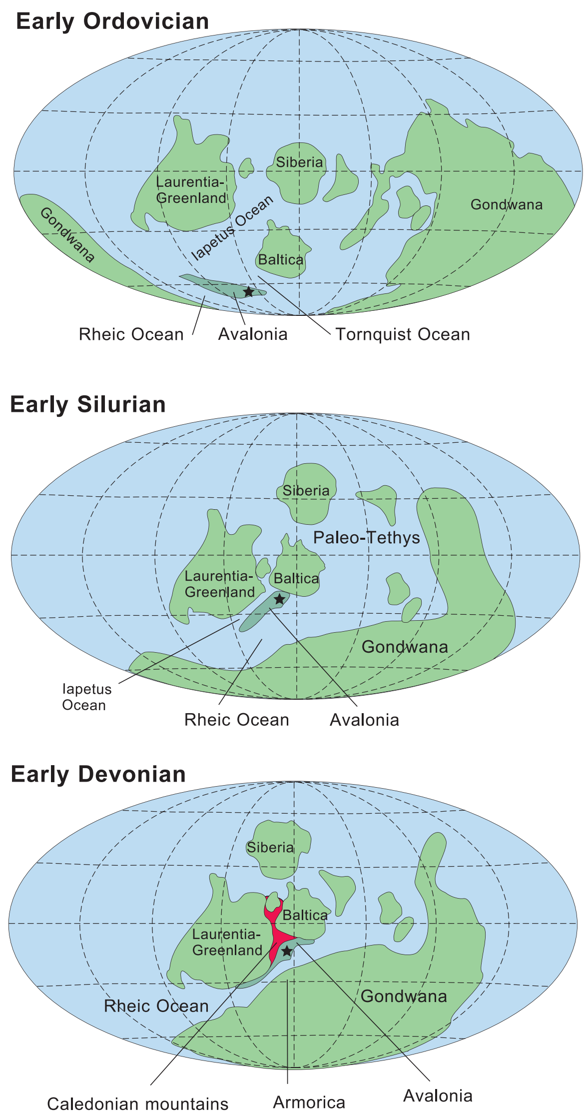

Fig. 4. Plate-tectonic reconstructions illustrating the northward drift of Avalonia and its collision with Baltica and Laurentia (after Scotese & McKerrow, 1990). The star indicates the paleo-position of the Netherlands.

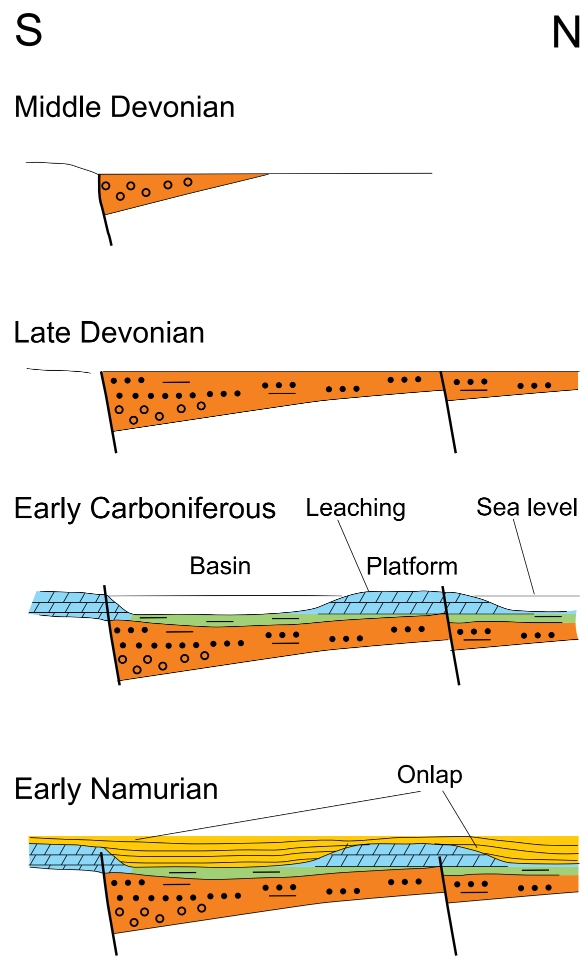

Fig. 5. Schematic model for the deposition of the Devonian and Lower Carboniferous in the area north of the Brabant Massif. Half-grabens controlled the sedimentation during the Middle Devonian to Early Carboniferous. Dinantian carbonate platforms reside on the footwall blocks, the hanging-wall blocks are characterized by basinal deposits. Exposure on the platforms led to karstification. The Namurian transgression progressively overstepped the platforms. After Bless et al. (1976), Muchez & Langenaeker (1993) and Fraser & Gawthorpe (1990). Legend as in Fig. 2.

ist below the Netherlands (Cameron & Ziegler, 1997), but poor seismic imaging of the Lower Carboniferous does not confirm this.

Alternatively to back-arc extension, Coward (1993) proposed his model of late Caledonian escape tectonics, with large-scale lateral expulsion of Baltica-Avalonia to the east, causing E-W extension in the southern North Sea area and deepening of the Rheic Ocean. In late Early Carboniferous times, differential subsidence in the North Sea area ceased and was replaced by more regional, thermal subsidence (Ziegler, 1990). A new stage in basin development began at the end of the Dinantian, when the Rheic Ocean closed, and the loading of the stacked nappes of the Variscan Mountains caused the development of the Silesian foreland basin in the North Sea area (Ziegler, 1990).

Lower Paleozoic, the Caledonian succession

In the Netherlands, the Caledonian succession has been encountered only in the offshore wells O18-1 and A17-1 and the onshore well Kortgene-1 (Fig. 6). The A17-1 well, situated on the Mid North Sea High, found a thick Upper Devonian succession overlying granite basement. This granite has been dated at 346 ± 7 Ma (Frost et al., 1981), but is believed to have been overprinted during the Early Carboniferous (Pharaoh et al., 1995). Recent U–Pb iso-

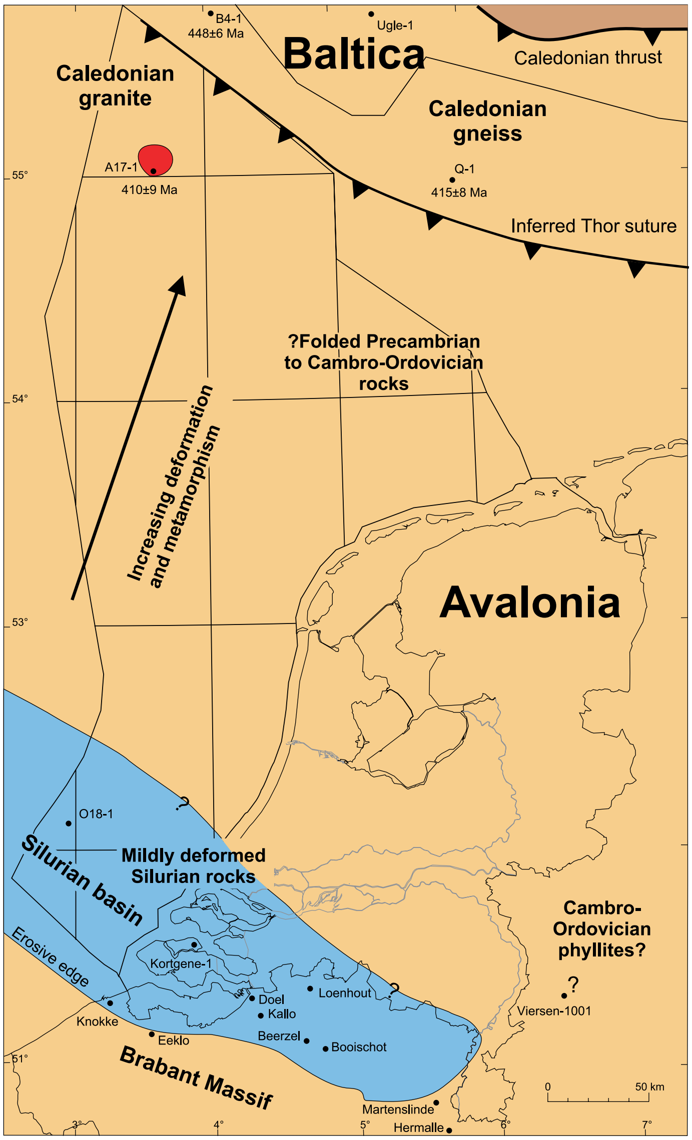

Fig. 6. Overview of observations on Caledonian deformation in the Netherlands. Wells shown reached pre-Devonian rocks.

tope studies (in progress by Axel Gerdes) now suggest an emplacement age of 410 ± 9 Ma (Tim Pharaoh, personal communication). The O18-1 well, situated on the northern flank of the Brabant Massif, penetrated a thick Lower Carboniferous and a thin Upper Devonian overlying dark, flat-lying, fine-grained turbidites of Silurian age. Frequent slumping is observed in these deposits. The well Kortgene-1 found a truncated Lower Carboniferous and thick Middle to Upper Devonian overlying dark-coloured Silurian rocks (Ludlow-Wenlock; NITG, 2003).

On the Krefeld High in Germany, Lower Ordovician rocks occur, as is documented by conglomerate clasts in the well Viersen-1001 (Ahrendt et al., 2001). Ordovician or Silurian rocks have further been encountered 75 km east of the Netherlands in the Soest-Erwitte-1 well (Clausen & Leuteritz, 1982), and Silurian rocks in east Anglia (Woodcock & Pharaoh, 1993). The German offshore well Q-1 encountered Precambrian muscovite-biotite augen gneiss, overprinted during the Caledonian orogeny; this gneiss is attributed to the Baltic Shield (Best et al., 1983; Pharaoh et al., 1995).

On the northern flank of the Brabant Massif in Belgium, wells encountered Cambrian (Martenslinde), Ordovician (Beerzel), and Silurian (Hermalle, Loenhout, Booischot, Doel and Kallo). Further to the south-west, several hundred drill holes reached the Lower Paleozoic, and many outcrops occur in river valleys on the crest of this massif (Legrand, 1968; De Vos et al., 1993; Debacker, 2001).

Three mega-sequences can be distinguished in the Caledonian succession of the Brabant Massif and England (Legrand, 1968; Verniers & Van Grootel, 1991; Woodcock, 1991; De Vos et al., 1993; Servais et al., 1993; Verniers et al., 2001, 2002): one from the Lower Cambrian to the Tremadoc (Lower Ordovician), a second from the Arenig to the Caradoc (Upper Ordovician), and a third from the Ashgill to the Upper Silurian (Fig. 2).

Mega-sequence 1 starts with coarse-grained terrigenous sediments of the Lower Cambrian, partly interpreted as turbiditic (Tubize formation), including quartzites, feldspar-rich metasandstones, siltstones and occasional phyllitic layers. The sediments become more fine-grained, pelagic and hemipelagic, towards the Upper Cambrian, and again turbiditic in the Tremadoc Chevlipont formation. No complete sections have been observed, and the total thickness of this mega-sequence exceeds 3700 m (Verniers et al., 2001, 2002).

After a long hiatus of at least 10 Ma, mega-sequence 2 was deposited first on a shallow shelf, dominated by fine siliciclastics, and finally, during the Caradoc, in a deeper environment with turbidites. The estimated total thickness of mega-sequence 2 is at least 850 m (Verniers et al., 2001, 2002). The presence of faults impedes observation of the stratigraphic contact with the overlying mega-sequence 3.

The sedimentation of mega-sequence 3 started on a shelf in the Late Ordovician and earliest Silurian. The sequence comprises fine siliciclastics, intercalated anoxic, dark, graptolite mudstones, and several volcanic and volcanoclastic horizons. In addition, porphyritic intrusive rocks of dacitic to rhyodacitic affinity are present with thicknesses of several hundreds of metres (e.g. Lessines, Ardooie). The hypabyssal porphyritic quartz-diorite intrusion of Quenast was described as a plug (André & Deutsch, 1984; André et al., 1986; André, 1991). During the Llandovery, the sedimentation became turbiditic with dominant fine siliciclastics, and a foreland basin developed; this is concluded from cumulative thickness curves (Woodcock & Pharaoh, 1993; Van Grootel et al., 1997; Debacker, 2001; Verniers et al., 2002). The formation of the foreland basin marks an orogenic inversion phase nearby, with crustal loading in the basin. The thickness of megasequence 3 is estimated at more than 3200 m in the outcrop area.

On a regional scale, the deformation of the Lower Paleozoic is relatively mild (Fig. 6). It increases in a north-eastern direction, together with the degree of metamorphism (Pharaoh et al., 1995).

Devonian

The Devonian in the Netherlands has been encountered in a limited number of wells. It non-conformably overlies the Caledonian substratum. The Lower Devonian is absent. The Dutch lithostratigraphy only defines Upper Devonian sediments (Van Adrichem Boogaert & Kouwe, 1994). The presence of Middle Devonian sediments, however, is likely on the basis of seismic data. The Devonian comprises the following lithostratigraphic units (Fig. 2):

- undefined Middle and lowermost Upper Devonian carbonates and siliciclastics;
- the Banjaard group (informal) in the southern Netherlands and north-eastern Belgium;
- the Old Red Group in the area of the Mid North Sea High.

In view of the large data gap between the areas, the relations between these units remain speculative. It is assumed, however, that the Old Red Group is present under a large part of the Netherlands, along with the extensions of the Middle Devonian carbonates (Ziegler, 1990).

Middle and lower Upper Devonian

Middle and lowermost Upper Devonian carbonates and siliciclastics have been encountered on the flanks of the Brabant Massif, on the Krefeld High, in the German Münsterland-1 and Q-1 wells, and in British wells west and north-west of the Dutch offshore (e.g. Auk and Argyll fields; Cameron, 1993). The rocks show a marked variation in composition (Fig. 7).

On the Krefeld High, up to 100 m of phyllite conglomerate with a Givetian age have been encountered in the Viersen-1001 well (Ribbert, 1998a). On the northern flank of the Brabant Massif, 400 m of conglomerate, of Givetian to Frasnian age, have been drilled in the Booischot well (Legrand, 1968; Kimpe et al., 1978; Streel & Loboziak, 1987). It is proposed that these conglomerates and the German Schwarzbachtal conglomerate were deposited in half-grabens close to active faults (Muchez & Langenaeker, 1993; Neumann-Mahlkau & Ribbert, 1998; Langenaeker, 2000; Fig. 5). They are covered by a Frasnian succession of reefal limestones, dolomites and claystones which has a thickness of some 400 m and an open-marine shelf facies similar to that of the type Frasnian in the Dinant Synclinorium (Ribbert, 1998b). This succession was also encountered in the Loenhout well and to the east of the

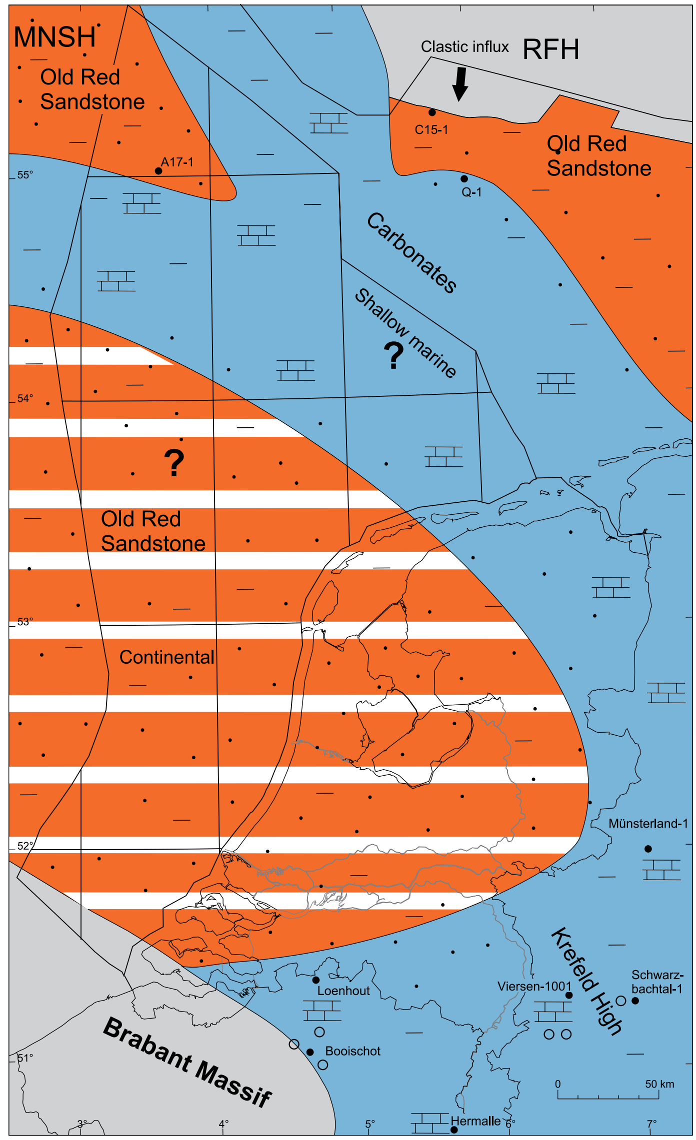

Fig. 7. Paleogeographic sketch map of the late Middle to early Late Devonian (Givetian-Frasnian). The brownish colour shows the extent of the continental Old Red Sandstone; in the hatched area it is assumed to be thin or even absent. Blue outlines the presumed occurrence of Middle Devonian carbonates (after Ziegler, 1990; Cameron et al., 1992; Kockel, 1998; NITG, 1998). Lithological symbols as in Fig. 2. RFH: Ringkøbing-Fyn High; MNSH: Mid North Sea High. Wells shown reached the mapped interval.

Brabant Massif on the Visé Block (Kimpe et al., 1978; Bless et al., 1981b; Poty, 1982). The sedimentation gradually expanded over the massif towards the late Frasnian sea-level highstand (Johnson et al., 1986), when northern biofacies crossed the crest of the massif (Conil et al., 1981), and the largely siliciclastic deposition of the Banjaard group began.

The carbonates around the Mid North Sea High have been defined in the UK lithostratigraphy as the Kyle

Group. This group was introduced by Cameron (1993) for Givetian limestones and overlying Frasnian mudstones and dolomites, with local anhydrite. It rests on Ordovician rocks in the Auk field (well 30/16-5) and is up to 170 m thick. Based on seismic data, the group is interpreted to be present on the southern flank of the Mid North Sea High, also in the Dutch sector (Cameron et al., 1992). In the A17-1 well, however, the carbonate has not been encountered, and younger, clastic, Devonian rocks overlie granitic basement. Based on seismic data (SNST 83 survey), the Devonian onlaps onto this granite. In the German Q-1 well, thin beds of marine Middle Devonian limestones are intercalated in the continental Old Red Sandstone (Best et al., 1983). It is assumed that the Kyle grades southwards into the Old Red Group. This latter group also overlies the Kyle Group.

The undrilled Dutch Middle and lower Upper Devonian is expected to contain in the southern Netherlands carbonate-dominated sequences, consisting of dolomites, biostromal limestones, nodular limestones and shales. These sequences, probably Frasnian in age, will be overlain by the Bollen claystone (Fig. 2). Along the northern margin of the Brabant Massif, they attain 33 m on the Heibaart dome, increasing to 90 m on the Visé Block, where they are strongly brecciated and karstified as a mogote (tower) karst. The towers are surrounded by solution collapse breccias and much younger Dinantian carbonates (Kimpe et al., 1978). Subvertical contacts between the tower karst and onlapping Viséan limestones are well-exposed near Visé in the Richelle quarries (Poty, 1982). The southern unit has not yet received a lithostratigraphic name in Belgium. It is suggested here to include it in a future formal description in the Banjaard group as the Heibaart Formation. It can be considered the equivalent of the northern Kyle Group.

An enigmatic Devonian succession occurs in the Winterswijk-1 well, containing 360 m of white-grey coloured sandstones. An arkosic sandstone with up to 25% plagioclase was cored in the lowermost part (NITG, 1998). Two large-scale coarsening-upward sequences were identified; in the upper part of the succession dolomite beds were encountered. Middle Devonian reefal limestones, as were encountered in the well Münsterland-1, 25 km to the east, are absent. There is limited biostratigraphic control, which leaves open several options for correlations (NITG, 1998; Karl-Heinz Ribbert, personal communication; Jacques Thorez, personal communication, favouring option i): i) the sandstones belong to the Condroz sandstone, and the Middle Devonian limestones occur stratigraphically deeper, ii) the sandstones present an unknown facies of the Middle Devonian below the limestones, or, iii) they are erosional products (or even part) of the epimetamorphic Lower Paleozoic basement.

Banjaard group

The Banjaard group has been introduced as an informal unit by Van Adrichem Boogaert & Kouwe (1994) for the late Middle Devonian to earliest Carboniferous, largely marine succession in the southern Netherlands. Equivalent successions occur in the adjacent parts of Germany and Belgium. The group comprises upper Frasnian and lower Famennian, dark, shaly mudstones with some thin, intercalated sandstone beds (informal name: Bollen claystone). These are overlain by the heterolithic Bosscheveld formation of Famennian to earliest Tournaisian (Strunian) age, comprising claystones and siltstones with intercalations of sandstones and nodular limestones with paleosols (Bless et al., 1981a). The thickness of the group reaches 300 to 700 m in the southern Netherlands, less on structural highs (NITG, 1999). In the eastern Netherlands it reaches 500 m.

To the south and east, part of the Bosscheveld formation grades into the shallow-marine Condroz sandstone, which has been encountered on the Krefeld High and further south in the Dinant Synclinorium in the Ardennes (Wolburg, 1970; Ribbert, 1998c). East of the Netherlands and also south of the Brabant Massif, the Condroz sandstone, deposited on the shelf around the massif from longshore currents, reaches a thickness of 800 m (Paproth et al., 1986; Thorez et al., 1988; Drozdzewski et al., 1998). The sandstone is more confined in age than the Bosscheveld formation, which contains several hiatuses that may eventually replace the totality of this formation along the basin margin.

Old Red Group

The Old Red Group comprises mainly fluvial, terrestrial sediments of Middle Devonian to Early Carboniferous age beneath the Dutch sector of the North Sea (Van Adrichem Boogaert & Kouwe, 1994). These sediments are broadly equivalent to the Upper Old Red Group of the UK southern North Sea. The group has not been penetrated completely in one single well, but based upon a compilation of well data a thickness of up to 1.5 km may be concluded on the Mid North Sea High. To the south, the group extends most likely under large parts of the Netherlands offshore and onshore, grading laterally into the largely marine Banjaard group (Figs 7, 8; Paproth et al., 1986).

Within the Old Red Group, three formations have been identified: the Patch, Buchan and Tayport (Cameron, 1993; Van Adrichem Boogaert & Kouwe, 1994). These formations are the equivalent of the UK Buchan and Tayport formations. In view of the limited data, only a brief lithological description will be presented below.

The *Patch Formation* (Givetian-Frasnian) comprises grey and green-grey silty claystones of lacustrine or floodplain origin, containing only a few thin sandstone beds. In the Netherlands, it has been encountered only in the A17-1 well, unconformably overlying Caledonian-aged granite. Its thickness is 300 m. In the UK, the Patch Formation forms the lower part of the Buchan Formation.

The *Buchan Formation* (Frasnian-Famennian) comprises a succession of white to reddish fluvial sandstones, conglomerates, and red to red-brown claystones and siltstones. In the eastern Mid North Sea High (well A17-1), acid extrusive volcanics (rhyolite) occur in the formation; they are a rather exceptional occurrence in the Old Red Group. Their radiometric age of 341 ± 30 Ma (Early Carboniferous) is of unknown reliability (Sissingh, 2004). Another, older (Middle Devonian), rhyolite occurrence has been reported from the Embla gas field (Norway, block 2/7; Marshall & Hewett, 2003). The formation has been encountered with a thickness of over 600 m in the north-western Dutch offshore and the UK.

The *Tayport Formation* (Famennian-Tournaisian) is composed of red to red-brown claystones and siltstones, with well-developed fluvial sandstone intercalations. Its total thickness is not known due to limited well control, but is at least 500 m in the Dutch sector, and 600 m in the UK sector.

Lower Carboniferous

The Dutch Lower Carboniferous or Dinantian (Tournaisian and Visean), is subdivided into two laterally equivalent groups (Van Adrichem Boogaert & Kouwe, 1994; Fig. 2):

– the Carboniferous Limestone Group, comprising partly silicified limestones and dolomites (traditionally known as Kohlenkalk);
– the Farne Group, comprising siliciclastics with intercalations of coals and limestone beds.

The nature of the transition between the groups is uncertain, owing to the large area not covered by sufficiently deep wells. Based upon released seismic data and literature (Maynard & Dunay, 1999), however, it is assumed that the carbonate facies extends north possibly as far as the Cleaver Bank and Schill Grund highs. Public-domain, seismic lines running across the Texel-IJsselmeer High show a high-amplitude reflector below the Upper Carboniferous, which is very similar in character to the reflector at the top of the Dinantian carbonates along the southern basin margin.

Based upon observations from the English onshore area, Cameron et al. (1992) assume that the thickness of the Lower Carboniferous may vary between 1000 and 2500 m in the southern North Sea. This might also apply to the adjacent Dutch areas. Scarce seismic data, however, indicate that the Lower Carboniferous on the highs, e.g. the Texel-IJsselmeer High, is much thinner, in the order of 200 m, possibly as a result of condensed sedimentation.

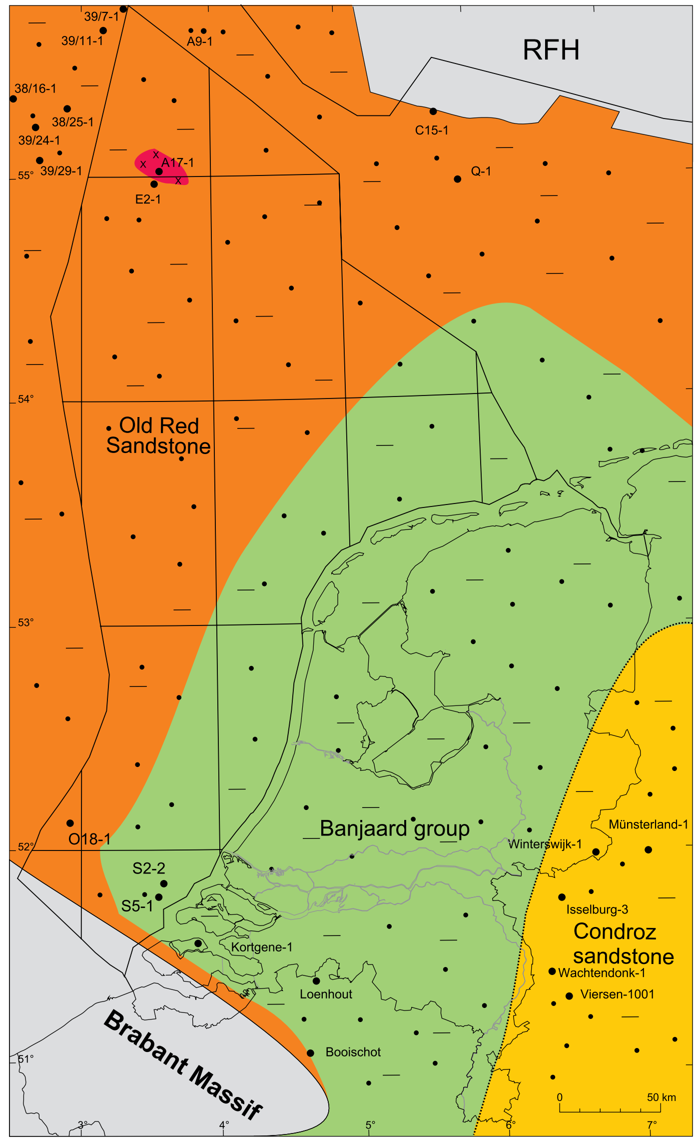

Fig. 8. Paleogeographic sketch map of the late Late Devonian (Famennian). The brownish area outlines deposits of continental Old Red Sandstone, the yellow area the marine Condroz sandstone, and the green area in between the largely marine Banjaard group (after Ziegler, 1990; Cameron et al., 1992; Paproth et al., 1986; Kockel, 1998; NITG, 1998). Acid volcanism occurred in the northernmost Dutch offshore (well A17-1). Lithological symbols as in Fig. 2; abbreviations as in Fig. 7. Wells shown reached the mapped interval.

Carboniferous Limestone Group

The Carboniferous Limestone Group contains light-grey, brown and black carbonates, with variable intercalations of cherts and claystones. The group is represented by the Zeeland Formation, and reaches a thickness of 750 to 1500 m in the south-western offshore (Cameron et al., 1992; Van Adrichem Boogaert & Kouwe, 1994). NW-SE striking syndepositional faults have been mapped in the

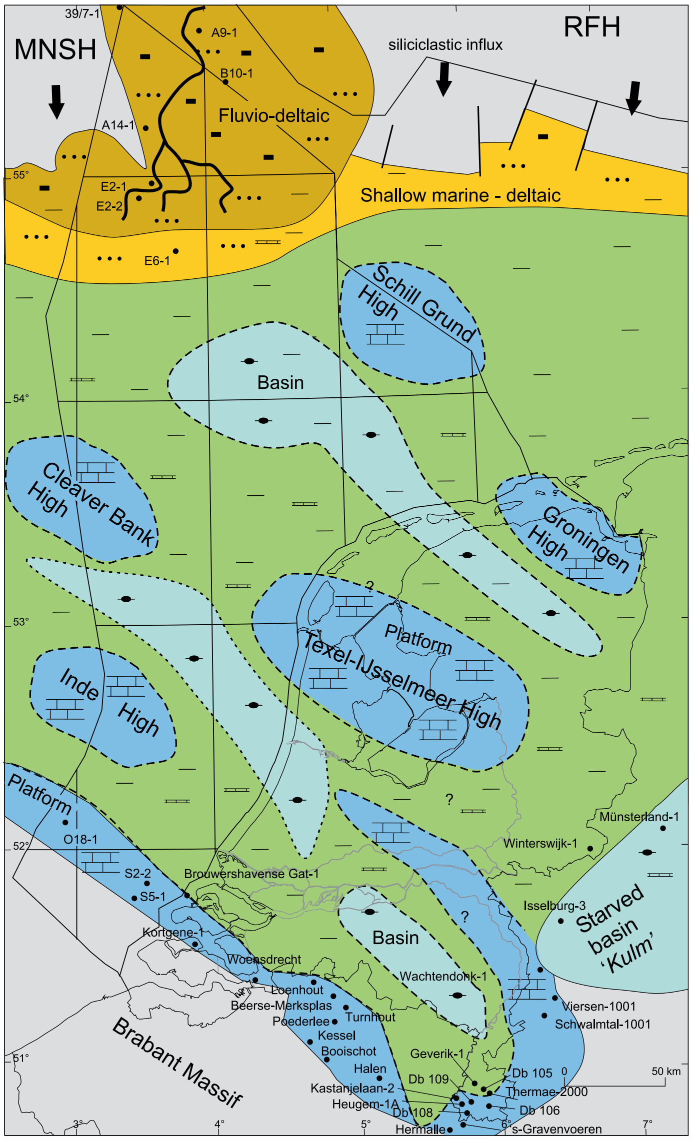

Fig. 9. Paleogeographic sketch map of the Early Carboniferous (after Bless et al., 1976; Lokhorst, 1998). Inferred carbonate platforms and intra-platform basins are shown. Lithological symbols as in Fig. 2; abbreviations as in Fig. 7. Wells shown reached the mapped interval.

UK (Tubb et al., 1986), and Vandenberghe (1984) identified E-W oriented faults north of the Brabant Massif. Also below the Netherlands syndepositional faults are assumed to be present (Quirk, 1993). No detailed mapping has been carried out, but it is likely that long-lived structural highs and fault zones, such as the Cleaver Bank, Schill Grund, Texel-IJsselmeer, Groningen and Inde highs and the Peel Block, formed shallow platform areas during the Dinantian. They were bordered by fault-bounded basins. Active faults are thought to have included the Peel Boundary, Hantum and Mid Netherlands fault zones. These uncertain assumptions are not included on the facies map (Fig. 9).

The Zeeland Formation is subdivided into three members, from bottom to top the Beveland, Schouwen and Goeree (Van Adrichem Boogaert & Kouwe, 1994). The Schouwen and Goeree members correspond respectively to the Belgian Kempen Limestone and Kessel formations (Paproth et al., 1983). Each member may reach a thickness of several hundreds of metres. This thickness most likely varies depending on the paleogeography. The *Beveland Member* (Tournaisian to early Visean, V1) is a sequence of medium- to dark-grey, coarse-crystalline dolomites with black residual organic matter (up tot 5.5 wt%, commonly concentrated in thin laminae or around recrystallized dolomite; Bless et al., 1981a) and locally well-developed intergranular to vuggy porosity. In places, limestone, claystone and siltstone intercalations occur. The *Schouwen Member* (middle to late Visean, V1-V3a) is composed of grey to brownish limestones. These limestones are mostly micritic, locally oolitic and abundantly fossiliferous, indicating depositional environments ranging from lagoonal or intertidal to shallow subtidal. Intergranular, bituminous material is often present. Thick breccia units, possibly of both dissolution and dynamic origin, occur locally within and on top of the member. Near the edges of the Brabant Massif, this member directly overlies the Caledonian basement. Reduced subsidence was accompanied here with the development of paleosols (silcretes, rootlet beds), thin coal seams and some siliciclastic influx. The *Goeree Member* (late Visean, V3) is a succession of grey to black limestones, which are thin- to thick-bedded, and often partly silicified. Black shales and cherts, and also thin tuffaceous beds occur in the upper part of the member. On structural highs, this member may be represented by microbial or cryptalgal reef mounds (Muchez et al., 1991; Mathes-Schmidt & Elfers, 1998).

Faults had a major effect on the facies distribution and reservoir potential of the group. They separated the basin in fault-bounded highs and lows. A typical example of a high is the Heibaart dome in northern Belgium (Kimpe et al., 1978; Vandenberghe, 1984). Sedimentation here was discontinuous, and subaerial exposure occurred prior to deposition of the overlying Silesian (Bless et al., 1976, 1980). Such exposure occurred mainly during the latest Visean and resulted in extensive leaching and karstification (Vandenberghe, 1984; Dreesen et al., 1987). Especially the upper Visean (V3) and lowermost Namurian are missing. Similar facies and leaching can be expected on footwall fault blocks of intra-basinal highs, such as the Schill Grund, Cleaver Bank, Inde, Texel-IJsselmeer and Groningen highs and the Peel Block. Lack of seismic data and poor seismic imaging, however, did not allow a proper delineation of these highs.

In the basins and on the hanging-wall blocks, sedimentation was more continuous and a thick succession of dark siliceous limestones (V3c) was deposited. This is documented by the Geverik-1 well (NITG, 1999).

The starved-basin or Kulm facies in the Carboniferous Limestone Group drilled in Germany comprises a strongly reduced, but stratigraphically complete succession (30–35 m) of thin-bedded, alternating, partly silicified limestones and dolomites, with some tephra layers (Münsterland-1, Isselburg-3; Fig. 9; Wolburg, 1963, 1970). Trilobites, crinoids and corals have been encountered (Wolburg, 1963). It is likely that this facies also occurs in the Netherlands in the basins between the carbonate platforms (Bless et al., 1976).

The matrix porosity of the Dinantian carbonates is only a few percent, except in the secondary dolomites of the Beveland Member. The fractured-reservoir porosity of the carbonates is controlled by karstification prior to the Namurian transgression. The longer the hiatus, the stronger the karstification. On the Heibaart dome, where the time gap between Dinantian and Namurian is 5 Ma, a super-permeable reservoir in Dinantian carbonates is used for the storage of natural gas. Namurian claystones seal this reservoir.

Farne Group

This group, deposited in a paralic to shallow-marine environment, comprises claystones and sandstones, with minor coal seams and a variable amount of limestone and dolomite beds. The carbonates increase rapidly towards the south. Three formations have been defined, from bottom to top: the Cementstone, Elleboog and Yoredale (Cameron, 1993; Van Adrichem Boogaert & Kouwe, 1994). The Cementstone and Yoredale formations are the equivalents of the formations defined in the UK, whereas the Elleboog Formation comprises both the Fell Sandstone and Scremerston formations. The group has not been penetrated entirely in one single well, but is assumed to reach a thickness between 1 and 2.5 km in the UK sector of the North Sea (Cameron et al., 1992). Strong variations in thickness, in the order of hundreds of metres, occur over syndepositional fault zones south of the Mid North Sea High (Maynard & Dunay, 1999).

The *Cementstone Formation* (Tournaisian) is composed of a cyclic alternation of carbonates, claystones and sandstones and minor coal seams. The carbonates occur as limestones, dolomitic limestones and dolomites. The thickness of the formation in wells is up to 400 m (e.g. UK well 44/2-1; Cameron, 1993).

The *Elleboog Formation* (early to middle Visean) comprises sandstones and claystones, with minor intercalations of coal and limestones. To the west, the amount of sandstones in the lower part increases, grading into the fluvial Fell Sandstone Formation of the UK southern North Sea (Cameron, 1993). The sandstone is massive, pale-grey and white, fine to medium or coarse-grained and poorly sorted. It contains sporadic beds of grey, blocky to subfissile, non-calcareous mudstone up to 10 m thick and occasional beds of dark-grey, pyritic, carbonaceous mudstone. Three sandstone units, each 80 to over a 100 m thick, have been identified in the UK offshore blocks 43 and 44 (Cameron, 1993). In a southward direction, the sandstones rapidly grade into mudstones with interbedded thin sandstones (e.g. E6-1). The sandstones are overlain by fine-grained deposits (UK Scremerston Formation), comprising alternations of sandstone, siltstone, mudstone and coal, with occasional thin dolomite or limestone beds. Their thickness is around 100 m.

The *Yoredale Formation* (late Visean, locally earliest Namurian) comprises a cyclic alternation of limestones, claystones, sandstones and rare coal seams. The number of limestone beds varies, and rapidly increases towards the south. In the Netherlands, a thickness of 175 m was encountered in wells (e.g. A14-1, E2-1, E6-1), whereas in the UK sector, over 700 m were penetrated (41/24a-2; Cameron, 1993).

Paleogeographic development

The Cambrian to Early Carboniferous deposition took place in a wide variety of sedimentary basins. The rapid northward drift of the Netherlands, mainly during the Ordovician, was partially responsible for this variety (De Vos, 1998; Cocks, 2000). Reconstruction of the sedimentary basins during the pre-Devonian remains speculative for most of the area, whereas it is better documented for the Late Devonian and Early Carboniferous.

The Cambrian and Lower Ordovician deposits of mega-sequence 1 of the Brabant Massif are mostly marine and terrigenous, derived from the Gondwana continent, as witnessed by acritarchs and other fossils (Cocks & Fortey, 1982; Cocks et al., 1997; Servais & Fatka, 1997).

After the rifting off from Gondwana, sedimentation of mega-sequence 2 took place mainly on a shallow shelf, with limited erosion products being supplied inside the relatively small Avalonia microcontinent. At the end of this sedimentation, in the late Caradoc and Ashgill, the faunal affinity was with Baltica, and sediments started to arrive from that continent (Giese et al., 1994). Large-scale slumps in the Brabant Massif (Debacker et al., 2001) and the onset of turbiditic sedimentation in the mid Caradoc point to basin instability. At the time of the soft collision between Avalonia and Baltica, the Ardennes were uplifted and deformed during the Ardennian phase (Verniers et al., 2002), and subsequently eroded.

North of the Ardennes, a rapidly subsiding, Silurian foreland basin developed during the deposition of mega-sequence 3 over the Brabant Massif and its western prolongation in east Anglia (Fig. 6; Woodcock & Pharaoh, 1993; Pharaoh et al., 1995). Turbidity currents supplied sediments, probably from southern source areas (Verniers & Van Grootel, 1991). During the final, Acadian or Brabantian phase of the Caledonian orogeny in the Late Silurian and the Early Devonian, inversion of this basin formed the Anglo-Brabant deformation belt (Van Grootel et al., 1997). The Cambrian core of this belt was progressively compressed and deformed, while sedimentation in the foreland basin continued (Debacker, 2001; Verniers et al., 2002).

Northward, the deformation and metamorphism of the sediments increased (Pharaoh et al., 1995). In Late Silurian and Early Devonian times, granites intruded in the area of the Mid North Sea High, e.g. the Dogger Granite, and the granite encountered in well A17-1.

In the southern North Sea area, the Netherlands and probably also the Brabant Massif, back-arc rifting in Middle Devonian times triggered the development of a horst-and-graben topography (Ziegler, 1990). Some of the horsts are granite-cored (De Vos et al., 1993; Corfield & Gawthorpe, 1995). In the grabens, continental, coarse-grained sediments of the Old Red Group accumulated. These sediments originated from source areas in the north-west, and from the London-Brabant Massif. Those from the latter area were mainly deposited in the Rhenohercynian Basin, south of the massif (Kockel, 1998). Alternative tectonic models to the back-arc rifting have been proposed by Coward (1993) and Maynard et al. (1997), and involve large-scale oblique strike-slip movements along the plate boundaries to explain the N-S extension in the North Sea area.

Middle Devonian marine sediments in the southern North Sea were deposited in an embayment of the Rhenohercynian Basin, that developed by northward transgression along a series of grabens, with WNW-ESE and N-S trends. The latter trend would be adopted later by the Dutch Central Graben. The horst-and-graben topography is thought to have played an important role, on the one hand by effectively trapping the siliciclastic sediments, on the other by creating a system of basins along which the transgression could reach this far north. It should be realized that only the extreme northern and southern margins of the carbonates are known, around the Mid North Sea High (Kyle Group) and on the northern flank of the London-Brabant Massif and the Krefeld High (Heibaart formation) respectively; they may well extend below most of the north-east Netherlands onshore area (Fig. 7). Also, it is most likely that a wide variety of sediments were deposited in the fault-bounded basins, ranging from fully marine carbonates, via sabkha sediments to lagoonal and finally continental clastics. On intra-basinal highs, such as the A17 granite, no carbonates were deposited.

During the Famennian, fluvial systems of the Old Red Sandstone built out over most of the Netherlands, and the shallow-marine Condroz sandstone was deposited in the eastern Netherlands and adjacent parts of Germany and Belgium (Fig. 8). These sands came from north-eastern source areas (Thorez et al., 1988). Rhyolite volcanism occurred in the Dutch and Norwegian parts of the North Sea.

Renewed transgressions during the Tournaisian and Visean pushed back the clastic influx to the Mid North Sea High area, whereas sediment supply from the Brabant Massif had ceased. An extensive carbonate platform was established in the southern Netherlands and adjacent areas (Fig. 9). Differential subsidence of the horst-and-graben topography continued, strongly influencing the facies distribution. Fluvial systems dominated the Mid North Sea High, and lacustrine to shallow marine deposition took place on the hanging-wall blocks south of the high (Maynard & Dunay, 1999). The clastics originated from northern source areas; they represent a delta system with regular marine flooding, responsible for the intercalated carbonate layers. The horst-and-graben topography effectively caught the siliciclastic deposits south of the high, and shielded the carbonate realm from clastic influx.

Further south, widespread carbonate deposition took place in the Netherlands, and the inferred depositional model includes a series of fault-bounded carbonate platforms and basins. Sea-level lowering during, and at the end of the Visean resulted in karstification on these platforms (Fig. 5). In the basins, sedimentation of fine-grained carbonates was continuous, and at times euxinic conditions prevailed.

Crustal sagging in front of the progressing Variscan orogeny buried the pre-Silesian relief under a thick coal-bearing molasse sequence. During the Namurian the previously emergent carbonate platform gradually became flooded.

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Silesian

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D.G. den Hartog Jager

Abstract

A thick Silesian sequence was deposited throughout the Netherlands and its adjacent North Sea area. Despite intense erosion during the Late Silesian and Early Permian much of it remains beneath the Permian and post-Permian successions. The preserved thickness locally exceeds 5500 m. The sequence comprises mainly siliciclastic sediments, deposited in a foredeep basin that formed as a result of the collision between the northward drifting Gondwana and Laurussia continents. The Variscan orogenic belt, the centre of this collision, formed the sediment source for the southern part of the basin. Sediments in the north were sourced from the Caledonian Mountains. The lithology ranges from claystone, siltstone and sandstone up to conglomerate. The units were first deposited in a lacustrine, occasionally marine basin bordered by deltas, alluvial plains and fans, and finally in a largely fluvial, red-bed environment. Peat was deposited from the Namurian C until the Early Westphalian D in deltaic and floodplain environments. The coal, which formed when this peat became deeply buried, has been mined in the southern Netherlands from medieval times until 1974. Combustible gas (mainly methane), originating from Silesian coal, was an early exploration target of the petroleum industry. Since the discovery in 1951 of gas in Silesian reservoirs in the Coevorden field, numerous other discoveries have been made in the east and north-east Netherlands from the 1950s onward, and in the offshore from 1974 until now.

Keywords: Netherlands, Pennsylvanian, stratigraphy, paleogeography, tectonics, Variscan, coal, hydrocarbon resources, igneous rocks

Introduction

In the Netherlands the Silesian, i.e. Late Carboniferous deposits are thicker and more widespread than those of any other Phanerozoic period. They are present below more than 95% of the country and its adjacent North Sea area, and even after major compaction, the package can be up to 5.5 km thick. This means that the Silesian is often thicker than all overlying sediments together (Fig. 1).

The Carboniferous of north-west Europe has been the subject of geological studies since the 18th century. The early interest may be explained by the excellent outcrops in the region, even within the Netherlands, and the presence of coal. For example, the landmark overview article on the Paleozoic of the Netherlands by Thiadens (1963) contains more than 20 pages on the Upper Carboniferous and just one page on the Rotliegend.

Both in Britain and in the Netherlands, the Carboniferous was among the earliest targets for hydrocarbon exploration. Coevorden, the first commercial gas field in north-west Europe, named after the nearby town which already had a gas-distribution network in 1951, produces from both the Carboniferous and the Zechstein. Of its initial reserves, 33×10^9 m^3 of gas are in Westphalian reservoirs, first appraised by well Coevorden-3 in 1951. Offshore, the first Carboniferous gas was discovered in 1974 in well K4-1D. A major breakthrough followed in the period 1984–1986 with the discovery of the D15-FA field, and of Schooner, Ketch, Murdoch and Caister in nearby UK quadrant 44. Unfortunately, poor reservoir quality and lack of lateral continuity often resulted in disappointing field developments.

Despite many decades of research, numerous aspects of the Silesian remain elusive. The scarcity of clear marker beds in outcrops and wells, as well as dating uncertainties have long frustrated the setting up of an unambiguous regional stratigraphic framework. The thick clastic Silesian packages may appear monotonous on quick view, but detailed investigations reveal that rapid lateral facies changes are common. Publications of investigations on the Silesian of the former coal mines in southern Limburg are numerous. However, results of these studies are not always representative for the Carboniferous Basin elsewhere in the Netherlands. Publications outside the coal-mining area are rare because the Silesian is deeply buried and only penetrated by wells drilled by petroleum companies that tend to keep their findings to themselves. The authors of this chapter had access to, and participated in, investigations of wells primarily drilled by the Nederlandse Aardolie Maatschappij (NAM). The investigations were compiled in regional studies and, together with gas-production experience, have yielded new information that, as this chapter attempts to summarise, adds much to the knowledge gathered in the coal mines.

Structural setting

During the Silesian, the northward drifting Gondwana continent caught up with the Laurussia continent. The collision of the African part of Gondwana with the European part of Laurussia took place near the equator and resulted in the formation of the Variscan orogenic belt (Fig. 2). The Variscan Mountains formed the main sediment supply to

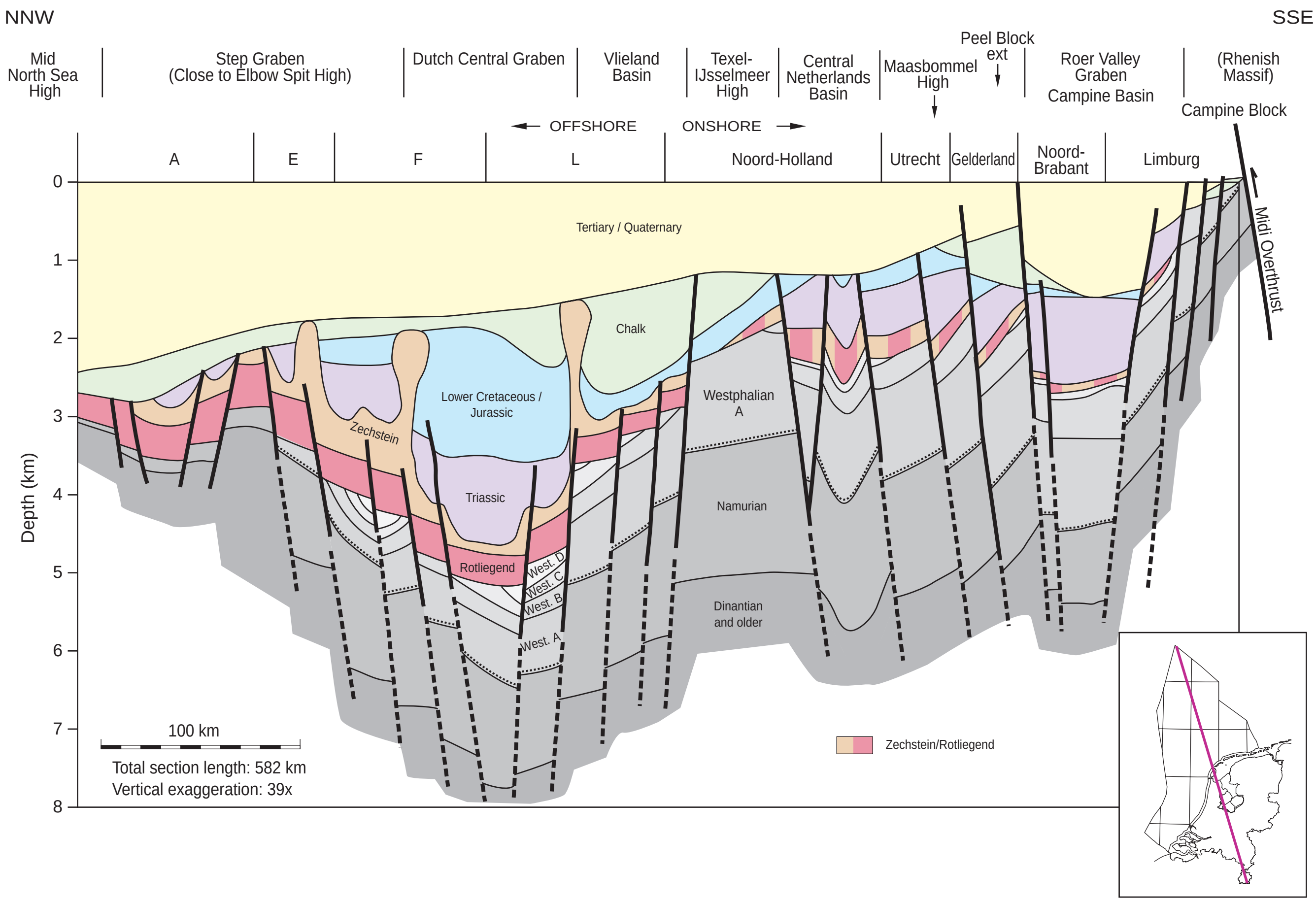

Fig. 1. Longest possible geological cross-section through the Netherlands on- and offshore, showing the overall structure of the relatively thick Silesian (Namurian and Westphalian) and its overburden. The Stephanian is not expected to occur in this section (cf. Fig. 13). The section also shows locations of structural elements mentioned in the text.

the foredeep basin north of this belt. Another supply originated from the 'passive' Caledonian hinterland north of the basin.

In the early phase of the collision, during the Namurian, rapid and uniform flexural subsidence took place throughout the Carboniferous Basin, with progressive onlap of siliciclastic sediments over the Dinantian platform carbonates of the London-Brabant Massif and the Mid North Sea High. A thick package of Namurian sediments in the Rhenohercynian Zone (Drozdzewski, 1993) and contemporary igneous intrusions suggest the presence of subduction systems to the south of this zone (Ziegler, 1990). Three subduction zones, two south- and one north-dipping, associated with Variscan deformation-front sutures, can be recognised between the Rhenohercynian and the more southerly Moldanubian Zone (Ziegler et al., 2004). Within the Dutch part of the basin, Silesian extrusives are recorded from onshore well Steenwijkerwold-1 (Overijssel) and inferred from 'hard' seismic reflections at the southern flank of the Elbow Spit High in the northern offshore.

During the Westphalian, the Variscan deformation front continued to migrate northward and the Namurian sediment package of the Rhenohercynian Zone was deformed into major thrust and nappe complexes (Ziegler et al., 2004). During the early formation of these complexes in the Westphalian A and B, the foredeep basin to the north continued to subside. In Germany, inversion-related uplift occurred from the Late Westphalian C onward. This continued throughout the Stephanian with an increase in magnitude and culminated during the Asturian phase, when a change in stress regime caused transpressional movements along NW-SE striking faults, as well as block tilting and the formation of NW-SE trending anticlines. The impact of this phase in the North Sea Basin is difficult to assess, since its effects became obscured by Early Permian thermal uplift and the subsequent substantial erosion of Carboniferous sediments.

Stratigraphy

Chrono- and lithostratigraphy

The north-west European Carboniferous has traditionally been subdivided into a carbonate-dominated Lower Car-

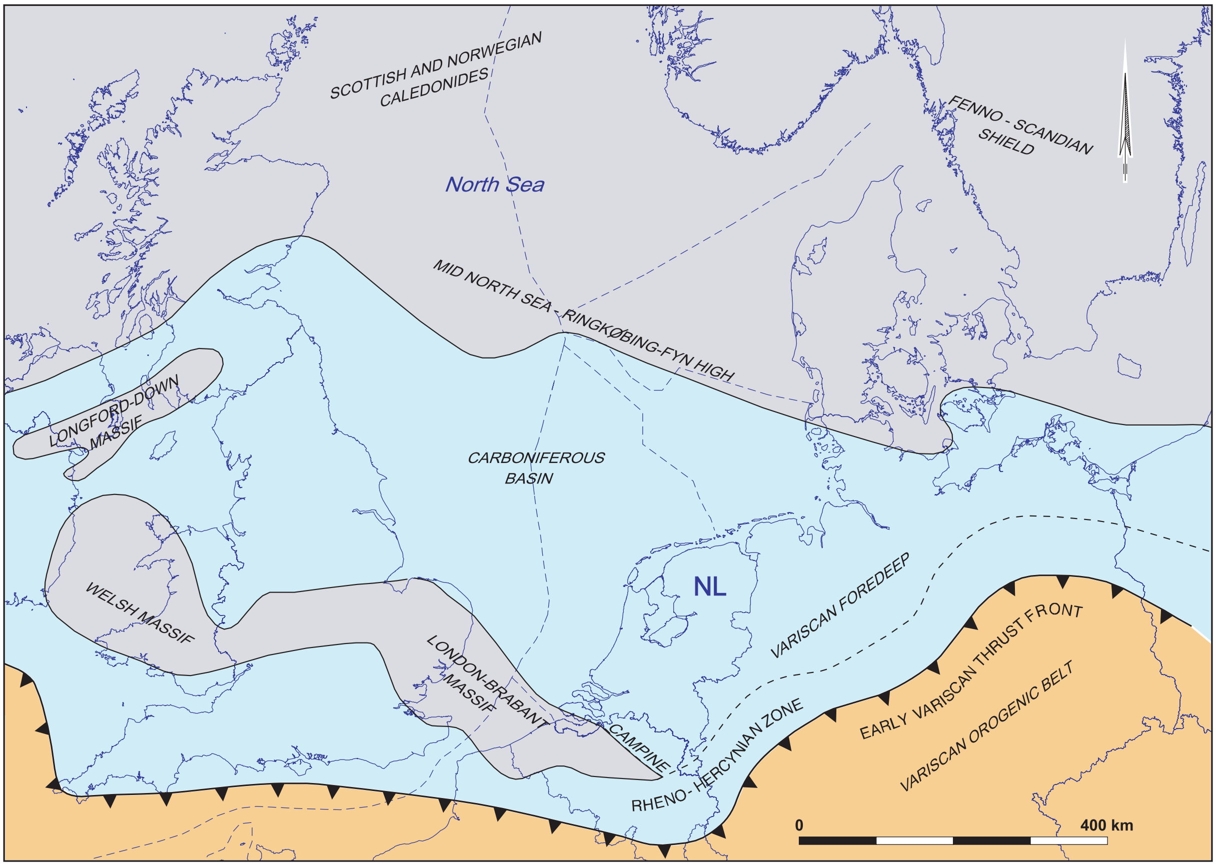

Fig. 2. Structural setting of the Silesian in north-west Europe (mainly after Ziegler, 1990).

boniferous and a siliciclastic Upper Carboniferous part. The lower part has been named the Dinantian and the upper part the Silesian subsystem. The actual Lower–Upper Carboniferous, i.e. Mississippian–Pennsylvanian chronostratigraphic boundary occurs at a stratigraphically higher level, close to the boundary between Namurian A and Namurian B (Fig. 3; Riley et al., 1994). The chronostratigraphic subdivision and correlation within the Silesian are based on macroflora, palynology (pollen and spores), 'marine bands' (calcareous claystone layers with marine, brackish or fresh-water faunas), weathered volcanic ash layers ('Tonsteine') and radiometric (Ar/Ar sanidine, U/Pb zircon and Pb/Pb zircon evaporation) ages of the ash layers (Menning et al., 2000).

The lithostratigraphic subdivision of the Silesian Limburg Group into subgroups is based on the recognition of four regressive cycles (Van Adrichem Boogaert & Kouwe, 1995). Based on a dominant or typical lithofacies characteristic, these subgroups are regionally subdivided into formations (Fig. 3). The boundaries of the subgroups and formations are diachronous. The lithology of these units ranges from claystone, siltstone and sandstone up to conglomerate (Fig. 4). The units were deposited in a lacustrine, occasionally marine basin, bordered by deltas, which developed into a basin with alluvial plains and fans. Peat was deposited from the Namurian C until the Early Westphalian D in deltaic and floodplain environments. A basin-wide peak in peat development took place during the Late Westphalian B and Early Westphalian C with the deposition of the Maurits and Westoe Coal formations.

Regional correlation

The stratigraphy, correlation and nomenclature of the Silesian have long been a matter of debate. As a result of successive congresses on Carboniferous stratigraphy in the Dutch coal-mining town of Heerlen between 1927 and 1955, some common ground was reached with the definition of north-west European series and stages that can be recognised basin-wide, i.e. Namurian A, B, C, Westphalian A, B, C, D and Stephanian. However, complex basin geometries, very thick sequences, the scarcity of marker fossils and the often endemic faunas and floras complicated stratigraphic correlations. Together with imperfect communication between scientists in countries in or around the Carboniferous Basin this led to an incomplete and often ambiguous stratigraphic nomenclature.

This situation markedly improved when the British Geological Survey published their standard lithostratigraphic

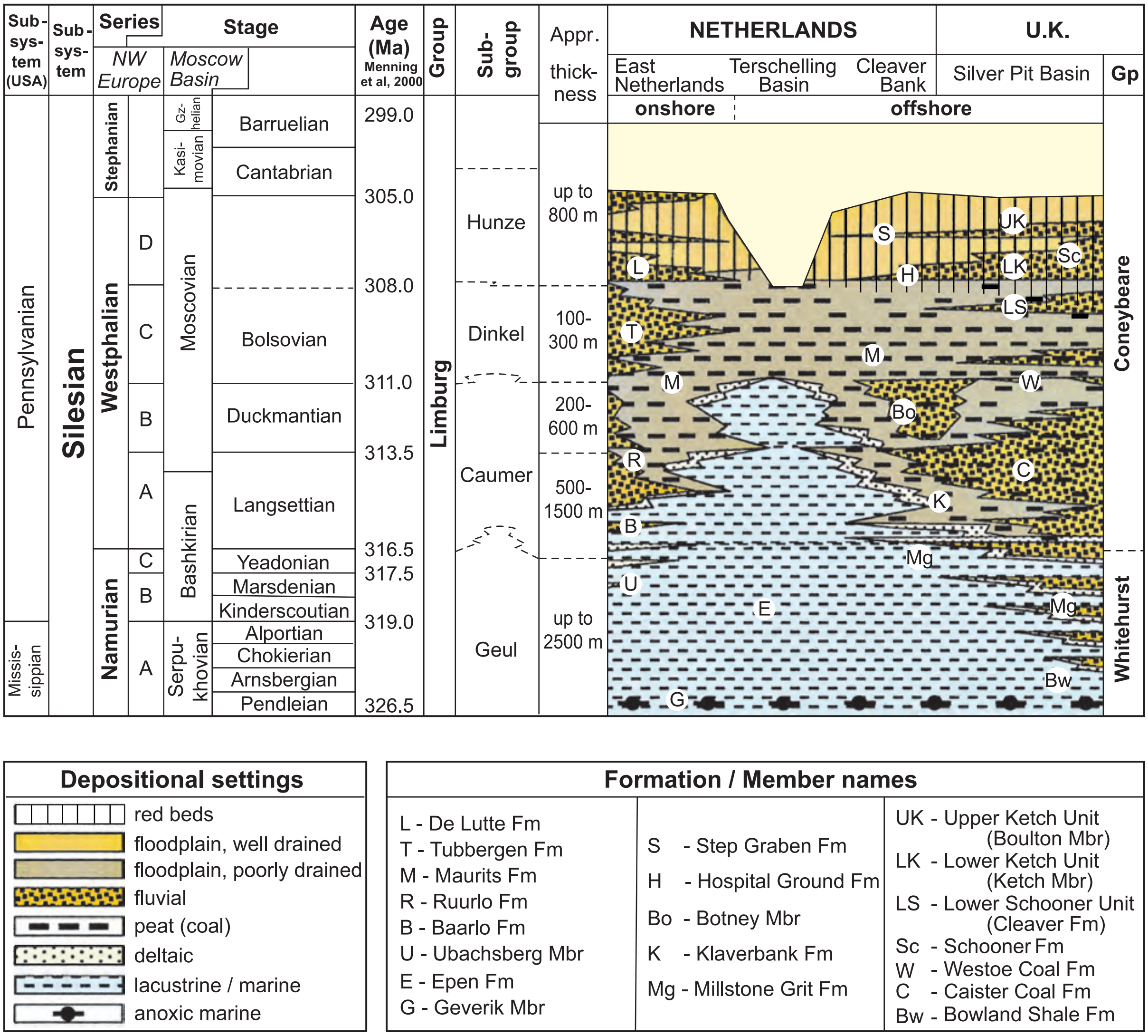

Fig. 3. Stratigraphic scheme for the Silesian of the Netherlands (after Van Adrichem Boogaert & Kouwe, 1995).

nomenclature of the UK southern North Sea (Cameron, 1993), followed by the Netherlands nomenclature compiled by the Rijks Geologische Dienst, the then Dutch Geological Survey (Van Adrichem Boogaert & Kouwe, 1995). In this chapter, simplified versions of both nomenclatures are compared and upgraded to honour additional information from recently released but often not published well data (Fig. 3).

Seismic character

The seismic character of the Silesian is very much associated with the presence or absence of coal within the sequence. The high impedance contrasts of the coal-to-shale and coal-to-sandstone contacts generate continuous seismic reflections, which are often grouped in easily recognisable 'reflective packages' (Evans et al., 1992; Quirk, 1993). The relatively low impedance contrasts between claystones and sandstones create only weak seismic reflectors in overall rather 'transparent' packages. Figure 5 displays an example from the Dutch offshore with a sandstone-dominated transparent package of the Klaverbank Formation (Westphalian A), a coal-dominated reflective package of the Maurits Formation (Westphalian B and Lower Westphalian C) and a sandstone and claystone-dominated transparent package of the Hospital Ground and Step Graben formations (Upper Westphalian C and Westphalian D).

Sedimentary cycles

The largest-scale cycle, which can be recognised, is the overall regressive cycle of the Limburg Group (Figs 3, 4). In sequence-stratigraphic terminology, this is a first-order cycle, which is associated with the subsidence and filling of the Variscan foredeep basin. The cycle starts with a succession of alternating marine shales, lacustrine shales and turbidites that develops along the edge of the basin into a succession of deltaic sediments toward the top of the Namurian. In the Westphalian, poorly drained fluvial and floodplain sediments, deposited around a lake of decreasing dimensions, are succeeded by well-drained fluvial, alluvial and floodplain sediments (Fig. 6). The cycle ends with well-drained alluvial and ephemeral lake sediments of the Stephanian.

At a smaller scale, four second-order megacycles can be recognised in the southern part of the basin. They coincide, more or less, with the Dutch lithostratigraphic subgroups. The thickness of the megacycles ranges from 100 to as much as 2500 m.

The Namurian megacycle Geul occurs on both the Dutch and British sides of the basin. The observed onlap of thick packages of largely marine and lacustrine sediments onto the London-Brabant Massif in the south and the Elbow Spit High in the north suggests that this cycle was formed as a result of rapid subsidence of the basin. The basin was subsequently filled by river systems encroaching from the south-east and north-west.

Although it is possible to correlate the Caumer and Dinkel megacycles of the Westphalian A, B and C on a regional scale, it is not possible to correlate them from one side of the basin to the other. This suggests that subsidence is not the main driving force controlling the deposition of these megacycles. In the southern part of the basin, the most likely driving force of both megacycles is possibly the alternation of overthrust loading and relaxation in the Variscan thrust belt to the south. During thrust phases, the topographic relief of this belt increases and generates a pulse of coarser sediments into the basin. During relaxation phases, subsequent decreases in relief as a result of further erosion reduce the amount and coarseness of the sediments.

In the northern part of the basin, the Westphalian A, B and C sequence does not consist of two but of three megacycles. They coincide, more or less, with the main part of the Dutch Klaverbank Formation, the Botney Member of the same formation and the Maurits Formation. The nature of the driving force of these three megacycles remains unclear.

The Hunze megacycle of the Westphalian D, and possibly lowermost Stephanian, occurs on the northern, as well as on the southern side of the basin. Contrary to the other cycles, this megacycle contains more sandstone at its base and more fine-grained sediments toward its top. The base of this megacycle has been amended in this chapter and now coincides with the bases of the De Lutte, Hospital Ground and Lower Ketch units. In view of the megacycle's drying-upward facies, observed on both sides of the basin, it can be concluded that the shift from a wet to a dry climate is the main controlling force for its deposition.

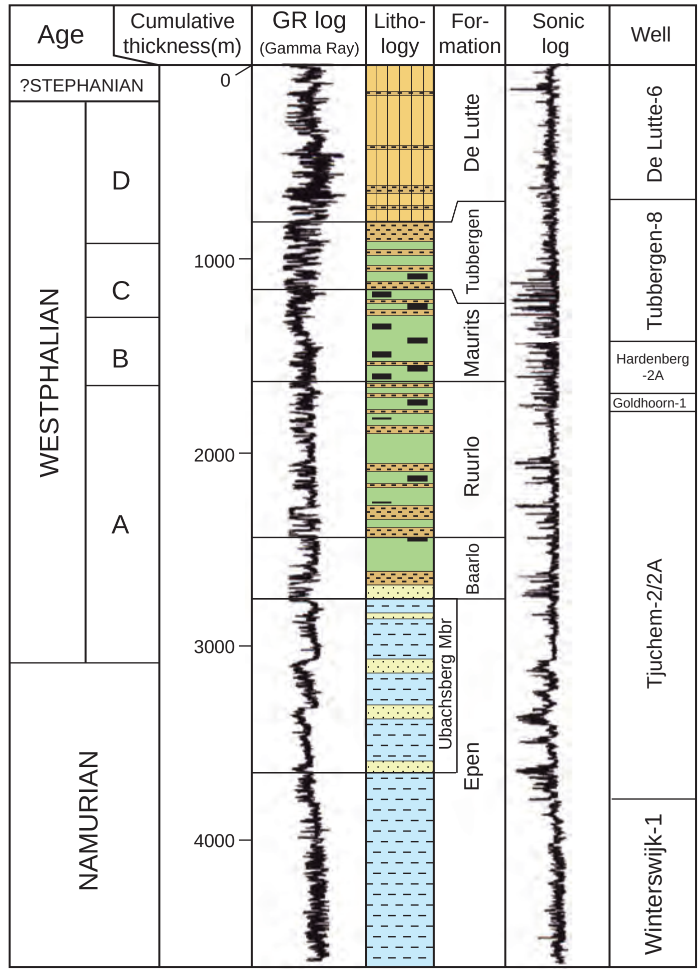

Fig. 4. Composite lithostratigraphic column through the Silesian of the north-eastern Netherlands, including composite sections of gamma-ray and sonic logs. For legend see Fig. 3.

The higher-order, i.e. smaller-scale, sedimentary cycles within the megacycles are probably associated with Milankovitch cyclicity. These cycles are governed by climate oscillations, reflecting periodic variations in the Earth's orbit around the sun (eccentricity), the variation in the angle between the Earth's axis and the ecliptic (obliquity), and the slow movement of the rotational axis of the Earth along a conical path (precession). During the Silesian, climate variations changed the intensity of water run-off and erosion in the hinterland and caused the sea level to fluctuate as a result of waxing and waning of the Gondwana icesheet. Large-scale glacio-eustatic cycles (third-order) are difficult to detect in the Carboniferous Basin. However, frequency analysis (Fourier transforms) on wireline logs from a 2200-m Namurian section Tjuchem-2 in the northern Netherlands disclosed a distinct sand influx pattern about every 250 m (Fig. 4). This approximately equates to the longest eccentricity periodicity of 2.1 Ma.

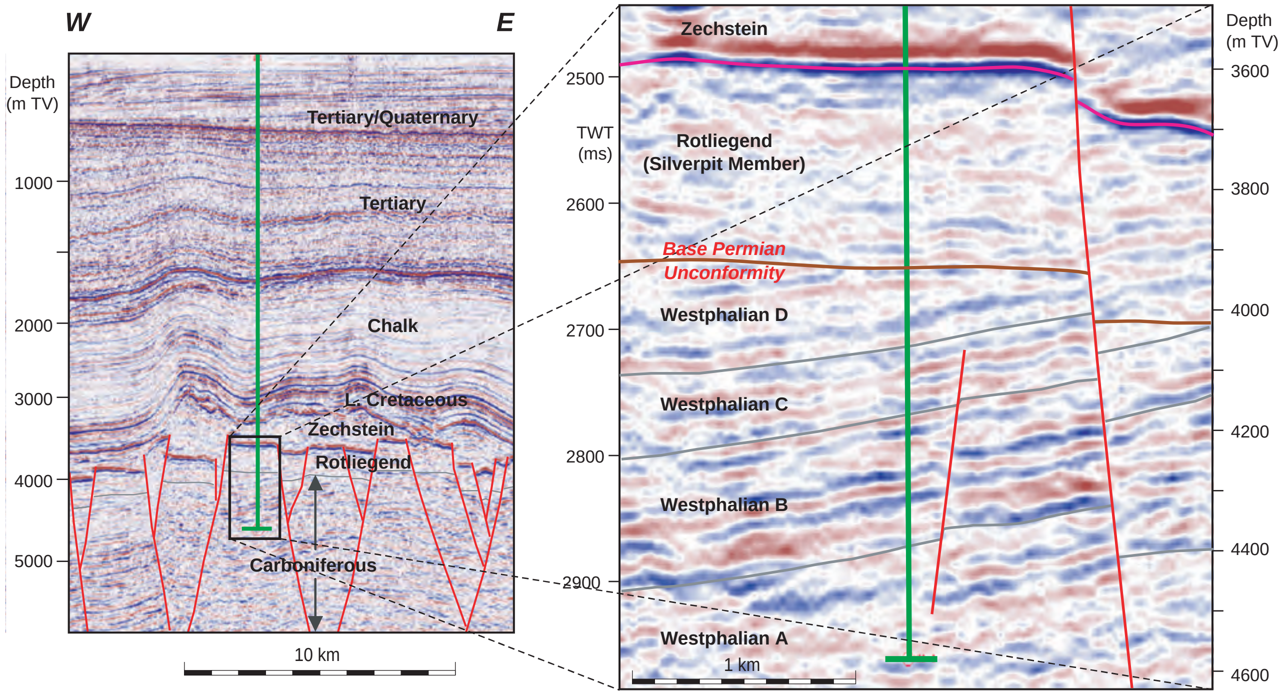

Fig. 5. Example of the Silesian as expressed on reflection seismic, Cleaver Bank High area. The green line represents well K2-2.

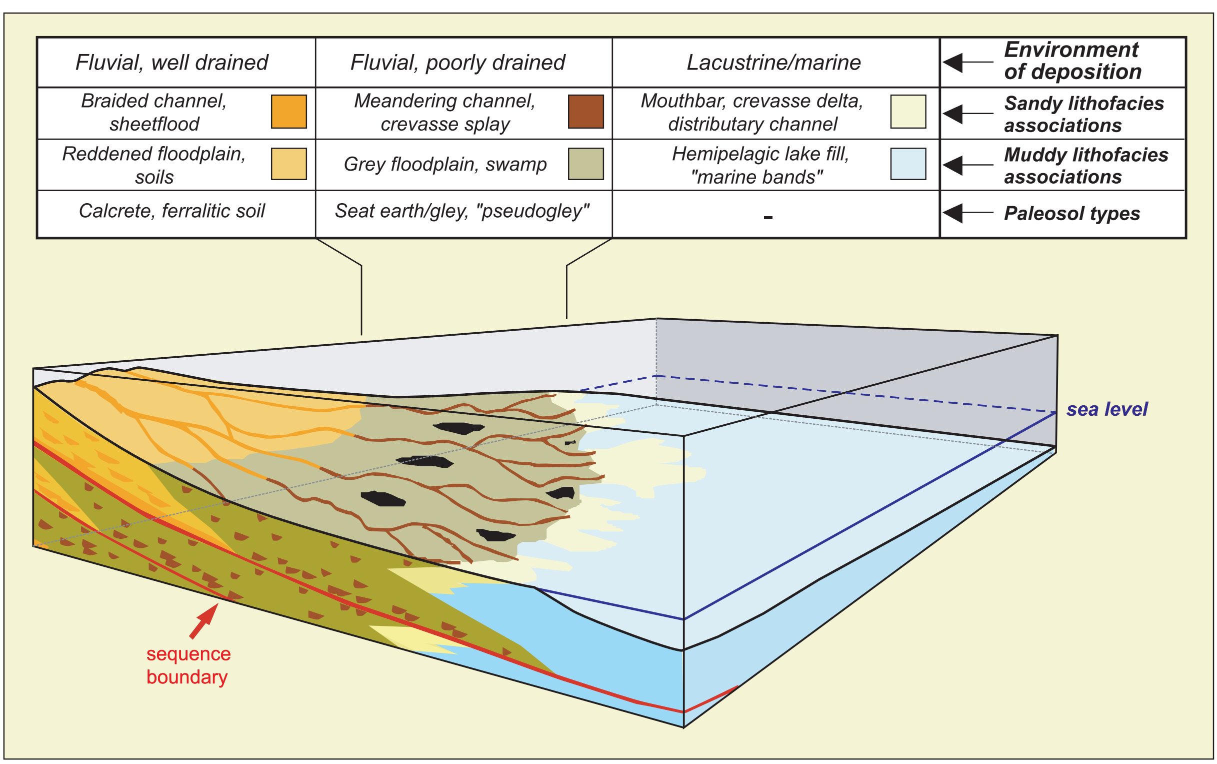

Fig. 6. Schematic distribution of depositional facies in the Silesian of the Netherlands. Black indicates peat.

Smaller-scale glacio-eustatic cycles of fourth or fifth order have been detected by applying frequency analysis on electric logs from the Westphalian D in well De Lutte-6 in the eastern Netherlands (Van Adrichem Boogaert & Kouwe, 1995: Fig. C.2). The analysis disclosed a distinct sand-shale alternation pattern every 100 m. Whether this pattern equates to the long (400 ka) or to the short (100 ka) eccentricity periodicity is not clear because of the inaccuracy in radiometric dating of the Westphalian D. In terms of thickness, the eccentricity periodicity in the underlying Westphalian and Namurian sections has been estimated to range between 40 and 100 m. This estimate agrees with sequence-stratigraphic studies in the Silesian of the Ruhr Basin in Germany, in which packages of alternating sand and shale, approximately 75 m thick, are tentatively correlated with fifth-order, short-eccentricity, (para-)sequences (Süss et al., 2000). At this scale, the 'classical' sequence-stratigraphic features, as described by Van Wagoner et al. (1988), can be recognised.

In most of the Dutch on- and offshore borehole penetrations of the Namurian A and B, low-stand shale deposits with intercalated turbidites are dominant. In the Namurian C, the turbiditic sequence passes into a sequence dominated by low-stand, prograding, deltaic wedges. Around the onset of the Westphalian, the depositional style in the Dutch on- and offshore gradually changed from a 'shelf model' in a predominantly low-stand setting, into a 'ramp model' in a transgressive and high-stand setting.

Sequence stratigraphy

Figures 6 and 7 depict the theoretical response of Westphalian sedimentation to a fall and subsequent rise of the base-level of erosion. In this model a fall in base-level will tend to cause net erosion. Existing rivers will start to incise; their deposits will become coarser, but will be eroded as long as the base-level keeps falling. This means that in the 'low-stand systems tract' and subsequent 'prograding wedge' phase, net deposition is only to be expected in the central part of the basin. When the base-level starts to rise again, first the additional accommodation space along the coastline will be filled, and subsequently fluvial deposits will be preserved as the river profile starts to backfill. The first deposits will be relatively coarse. Continuing base-level rise will eventually cause flattening of the topography. The erosion intensity will gradually decrease, and finally an overall fining-upward sequence will result.

During the 'maximum flooding' phase, the amount of terrestrial erosion will be minimal. Around this time, hy-

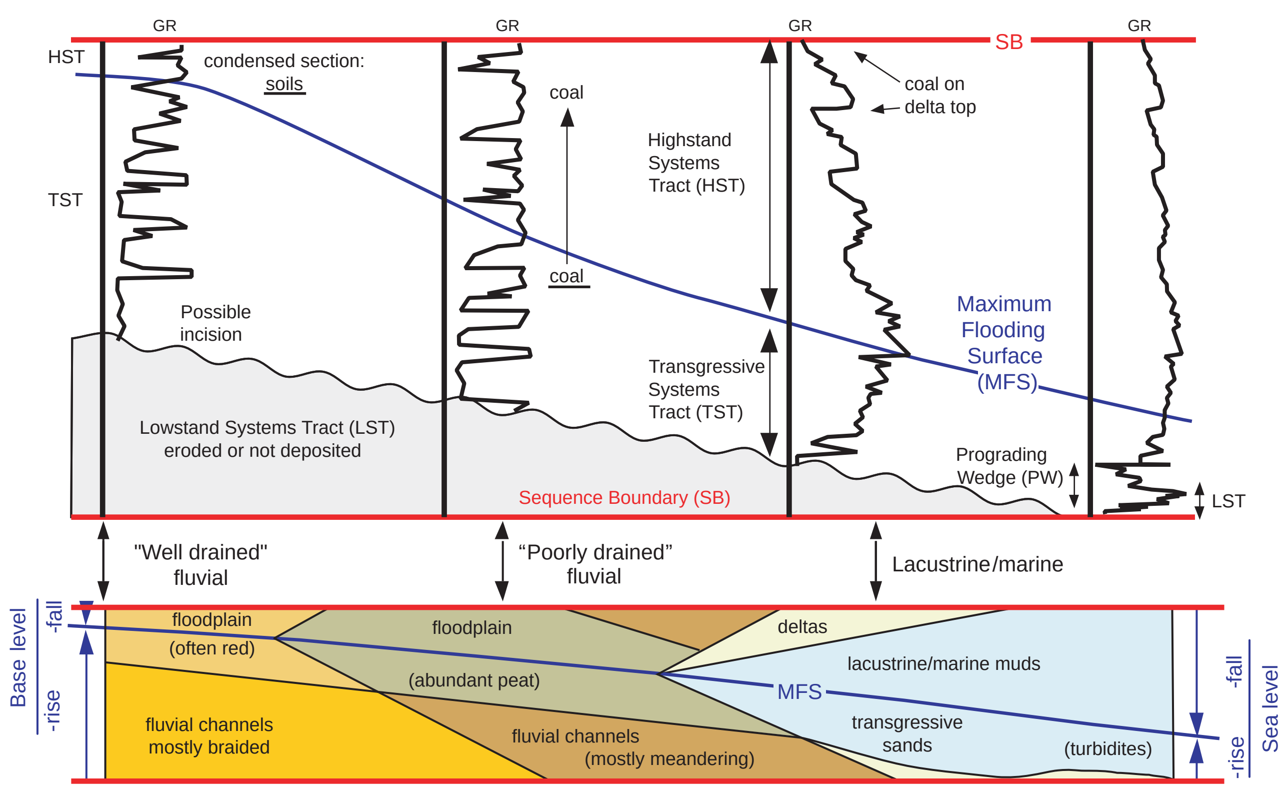

Fig. 7. Sequence-stratigraphic model showing conceptual gamma-ray logs (GR; top) and schematic facies distribution (bottom) for typical Silesian sequences.

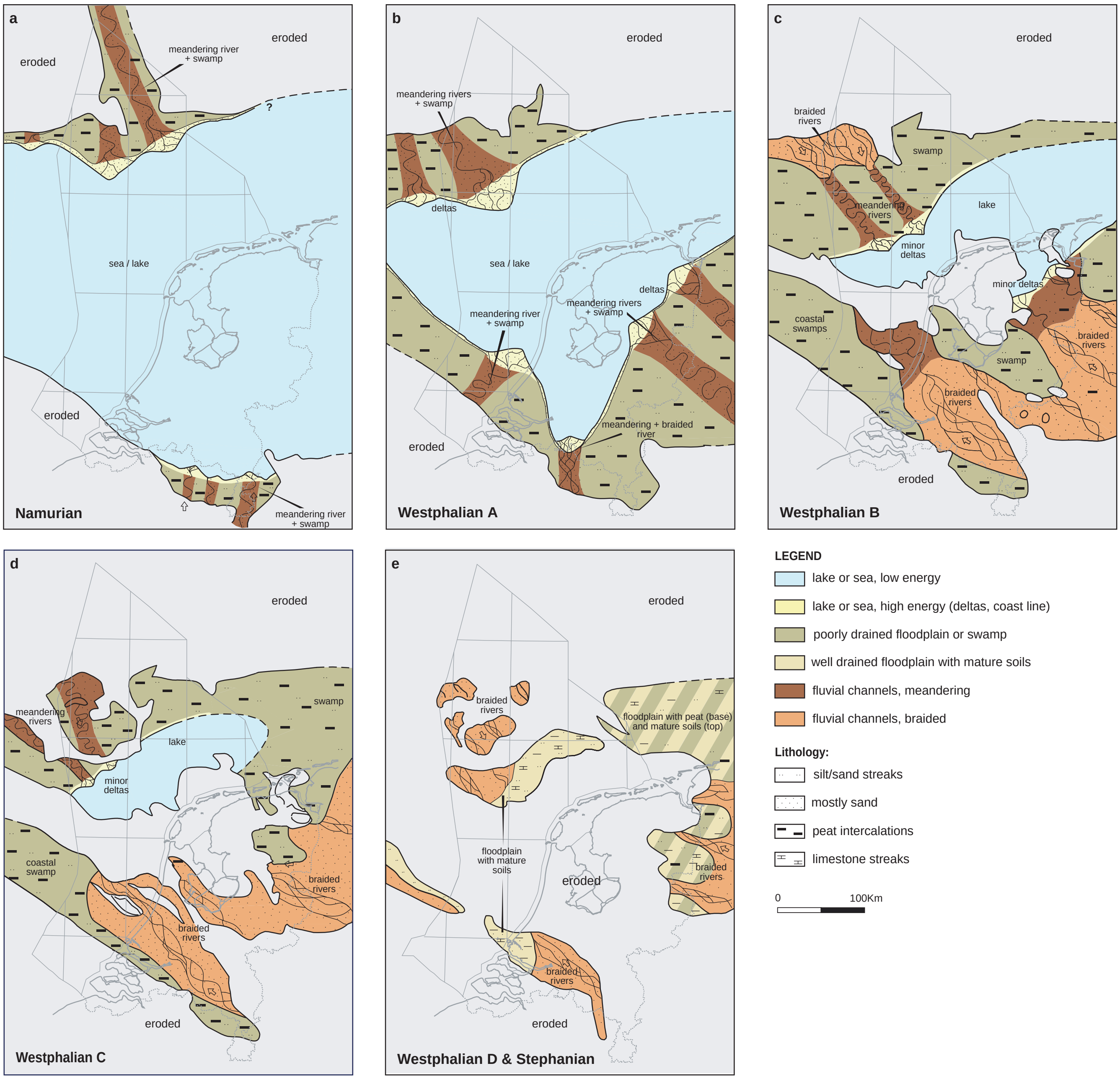

Fig. 8. Facies distribution maps. The indication 'eroded' refers to post-Silesian erosion (cf. Figs 9, 13). Westphalian D and Stephanian sedimentation took place in a red-bed setting (cf. Fig. 3).

drocarbon source rocks are likely to develop: peat in the poorly drained fluvial setting, where the vegetation profits from a high water table and a minimum of 'disturbing' sediment supply, and organic-rich clays in the condensed sequences within the lacustrine or marine setting. Westphalian maximum flooding surfaces are commonly recognised as 'marine bands', i.e. thin, argillaceous layers rich in marine and sometimes lacustrine faunas that can be excellent marker beds, e.g. the Aegir marine band at the base of the Westphalian C (Thiadens, 1963; Van de Laar & Fermont, 1990).

The fact that many of these decimetre-scale marker bands can be correlated over distances of several hundreds of kilometres illustrates how flat the basin was, at least during Westphalian A, B and C times. During deposition of the 'high-stand systems tract', the relative rise of the base-level slows down. The first material that is eroded upstream on the alluvial plain will be deposited downstream on the same plain. While the alluvial system will prograde over the downstream area, the coarser sand layers tend to thicken toward the top of the tract. Near the lake, small

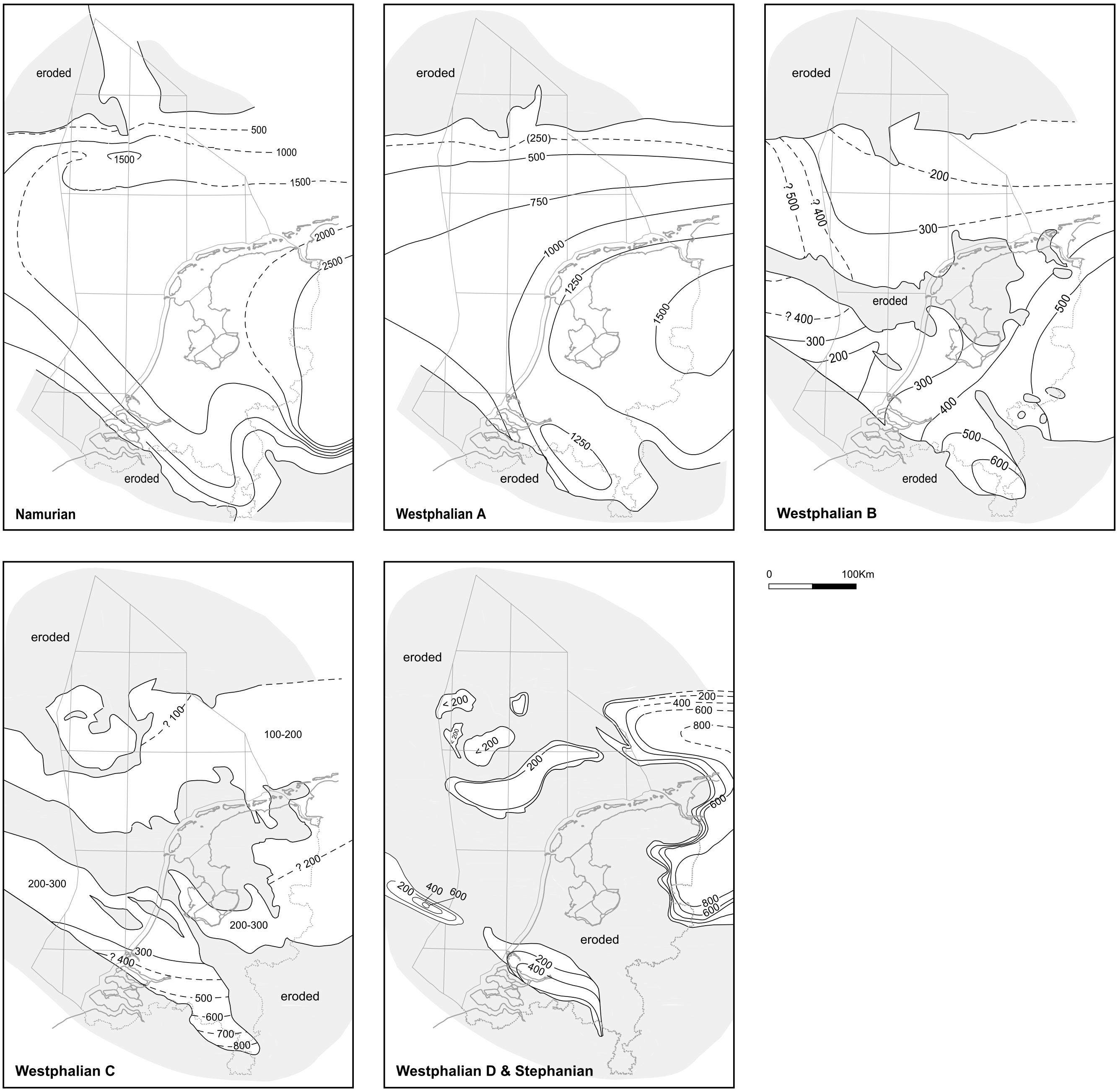

Fig. 9. Maps showing isopachs (in metres) for Silesian stages outside the areas of complete erosion. For Namurian to Westphalian C, original isopachs have been reconstructed. For Westphalian D and Stephanian, present-day isopachs are shown.

deltas may develop, which gradually grow bigger when the erosion intensifies. Where, in a fluvial setting, the net sedimentation of fine-grained material exceeds subsequent erosion, a 'type II sequence boundary' may develop. A condensed section with development of paleosols will mark this boundary.

The reasons why not every interval between two sequence boundaries shows exactly the predicted lithological succession are autocyclicity and parasequence formation. Autocyclicity simply implies that fluvial channels or deltas cannot be everywhere at the same time. They can switch position rapidly when sediment supply is increasing. The number and the sizes of channels are expected to increase in the lower part of a preserved fluvial sequence, but their actual distribution is a random process. At a particular location, a fluvial sequence may start with a substantial clay package. Parasequences are smaller, fifth or higher-order sequences, with a typical thickness of some 2 to 20 m. They may be related to the obliquity and precession of Milankovitch' cyclicity.

Sedimentary development

The facies maps of the main Silesian chronostratigrapic subdivisions in the areas not affected by post-Silesian erosion show the gradual transition from an initial largely marine and lacustrine area to a predominantly fluvial redbed environment (Figs 8a-e). Isopach maps of the same subdivisions indicate the considerable post-compaction thicknesses that are involved (Fig. 9).

Namurian

During the Namurian, the sedimentation of basinal clays and turbidites in the Carboniferous Basin was controlled by rapid subsidence. Initially, a marine environment prevailed, but gradually it gave way to fresh-water conditions. An organic-rich black shale, with a total thickness of some 25 m, occurs in the lowermost part of the Namurian section in well Geverik-1 (southern Limburg). The shale forms the lower part of the bituminous Geverik Member and has been classified as a type II source rock with a total organic carbon content of 8% (Van Balen et al., 2000). This Namurian A source rock, which developed in an anoxic marine environment, is thought to be present throughout the central and deeper part of the basin. Evidence from northern England indicates that these shales, generally referred to as Bowland Shale Formation, have generated oil, preserved in shows and some small fields (Kent, 1985). They may provide the charge model for new plays in the North Sea Basin as well (Gerling et al., 1999).

The fluvio-deltaic system which in Dinantian times shed its sediments southward from the Caledonian Highlands, continued to develop during the Namurian A in the north-western part of the basin (Fig. 8a). The distal equivalent of this system in the Dutch offshore is represented by the predominantly lacustrine to weakly marine sequence of claystones with turbidites of the Epen Formation. This sequence onlaps the Elbow Spit High (eastern Mid North Sea High) in the north and the London-Brabant Massif in the south. Its water depth is estimated at several hundreds of metres. During the Namurian B the northern fluvio-deltaic systems continued to prograde toward the basin centre, depositing delta-front sediments of the Millstone Grit Formation. During the Namurian C a similar system started to develop in the south, depositing the delta-front sediments of the Ubachsberg Member (Epen Fm), and the basin gradually shallowed. Toward the end of the Namurian, a persistent high sedimentation rate finally caught up with the basin subsidence and much of the lacustrine shallow-water basin was transformed into a basin in which deltaic and fluvial conditions prevailed. Intercalated marine bands indicate that marine incursions continued to occur during this phase of basin development.

In the UK offshore quadrant 43, Namurian stacked channel sands of the delta front, now the Millstone Grit Formation, form the reservoirs in the Trent and Cavendish gas fields, where they are partly developed in an incised valley-fill setting. In the Dutch sector, similar sands of the delta front, form the reservoir of two gas discoveries in the Namurian C to Lower Westphalian A section of the Klaverbank Formation in block E12. No incised valleys are known from this area.

In the onshore Netherlands, thick sections of Namurian shale with only minor amounts of sandstone were encountered in several wells, for example Tjuchem-2 (Groningen; Fig. 4) and Nagele-1. Further south, more sandstone, including sandstones of the already mentioned Ubachsberg Member, occurs in the Namurian of the Peel Block (Peelcommissie, 1963) and the Campine Basin (Langenaeker & Dusar, 1992).

Westphalian

During the Westphalian, a high sedimentation rate balanced the basin subsidence. The lake area in the basin centre gradually decreased. Along the margins, deltaic shallow-water sedimentation in the Westphalian A was replaced during the Westphalian B and C by swamp-dominated sedimentation with fluvial systems. During the Westphalian D, a climate shift from tropical to semiarid, combined with uplift tectonics of the encroaching Variscan front, replaced the wet, swamp-dominated environment by a dryer, floodplain-dominated setting with fluvial, ephemeral-flood and fanglomerate systems.

During the *Westphalian A*, the southern half of the basin was filled by delta systems prograding from the south (Langenaeker, 2000). At the same time, the northern half of the basin was filled from the north by a similar but smaller deltaic system (Fig. 8b). The Westphalian A consists of a series of coarsening upward cycles. The deposition of each cycle started during a brief lacustrine transgression or a marine incursion, after which a period of delta progradation followed. Regional paleogeographic maps suggest that the marine incursions came from the east, where the eastern extension of the Variscan foredeep probably connected with the Ural Ocean (Ziegler, 1988). In the upper part of a cycle, distributary- channel sandstones are intercalated with fines from the delta plain and with coal seams which originated in the coastal and delta-plain swamps. Westphalian A channel sandstones form the reservoir for the offshore gas discovery in Q13-3.

During the *Westphalian B*, fluvial sedimentation became more important (Fig. 8c). Around the shrunken central lake a succession of coastal and floodplain fines developed with frequent intercalations of peat and occasional intercalations of fine to coarse-grained and argillaceous sands. Meandering river systems deposited fining and coarsening upward sequences of up to 90 m thick. In the northern offshore a local but distinct fluvial system developed. It gave rise to the coarse-grained stacked channel sandstones, up to 50 m thick, of the Botney Member. These sandstones form the reservoirs in gas fields and gas discoveries in blocks D12, D15, E10, J3, K2 and K5. In UK quadrant 44, similar fluvial sandstones from a more westerly fairway constitute the reservoirs in the Tyne, McAdam, Caister and Murdoch gas fields.

During the Late Westphalian B and Early Westphalian C, the Carboniferous Basin was dominated by floodplains and swamps with extensive peat development (Maurits and Westoe Coal formations) and relatively small meandering river systems.

During the *Westphalian C*, a pronounced fluvial system developed in the eastern Netherlands, forming the Tubbergen Formation (Figs 3, 8d). Stacked channel sandstones of this formation, up to 100 m thick, form good reservoirs in the gas fields of the Coevorden area. In the giant Groningen gas field, sandstones of the Tubbergen Formation form part of the reservoirs in a gas column extending downward from the Rotliegend. In the northern half of the basin, floodplain and lacustrine claystones occasionally intercalate with relatively fine sandstone beds of meandering systems. Red-bed facies started to develop in the proximal part of a meandering depositional setting during the Late Westphalian C. In UK quadrant 44, the channel sandstones of this red-bed facies form good reservoirs in the Schooner and Ketch gas fields.

During the *Westphalian D*, red-bed facies became dominant throughout the basin (Figs 3, 8e). The corresponding interval consists mainly of reddish brown and greyish green, sandy and silty claystones. Mature paleosols are abundant. Intercalated greyish green and red fluvial sandstones form channels and sheets up to 30 m thick. Pebbly sandstones and conglomerate beds are frequently incorporated within intervals of fairly well sorted sandstone. The reservoirs in gas fields and gas discoveries in the offshore blocks D18, E13, E17, K2, K4 and K6 are Westphalian D sandstones.

Stephanian

Two wells in the Netherlands, De Lutte-6 in east Overijssel and F10-2 offshore, are reported to have penetrated a Stephanian section. In both wells a fluvial red-bed facies, similar to that of the Westphalian D, but with better-developed calcrete paleosols containing characteristic calcitic nodules, indicates a continuation of the drying-upward trend.

Reservoir-geological aspects

The possible absence and poor quality of reservoir rock form important risks for hydrocarbon prospecting in Silesian sediments. Predicting the presence of reservoir rock is difficult because of the lateral variability in sedimentological facies (Fig. 10), and predicting the quality of reservoir rock is difficult because of the common and complex diagenetic impairment. The main reservoir trend in the Silesian is determined by the overall progradational nature of the fluvio-lacustrine systems. As a consequence, the reservoir properties generally deteriorate from younger to older rocks and from basin fringe to basin centre (Figs 8, 11). Burial depth is the other important factor for the general prediction of reservoir quality (Figs 1, 12). An increase in depth, especially paleoburial depth, is usually associated with an increase in diagenetic impairment.

Multivariate analysis of petrographic data on thousands of core plugs from the Dutch Silesian has identified grain size as the main factor for porosity prediction. Larger grain size correlates with higher porosity. This means that facies distribution maps (cf. Fig. 8) are an important key to identification of Silesian prospectivity.

So far, most of the commercial gas in Silesian reservoirs has been found in fluvial channels of the Klaverbank Formation in the north-west offshore, and the Tubbergen Formation in the Coevorden-Groningen area. Reservoir sandstones vary from greyish white to pinkish red. The sand grains are fine to coarse, occasionally grading up to pebble and conglomerate size, moderate to poorly sorted and subangular to rounded (Van Adrichem Boogaert & Kouwe, 1995). Quartz-arenitic sandstones dominate the Klaverbank Formation, whereas the Tubbergen Formation is composed of a mix of lithic and quartz arenites (NAM, internal core description reports of cores from the D15-FA and Coevorden fields). Feldspar is a minor component of the reservoir sandstones. Both formations show major lateral variability but also a distinct decrease in reservoir quality in a sedimentologically distal direction over some 50 to 100 km. In both areas, the average reservoir porosity is around 9%, and permeabilities are around 1 to 2 milliDarcies only. However, the best reservoir intervals can still have a porosity exceeding 20% and a permeability exceeding 100 milliDarcies. The thickness of individual reservoir bodies ranges from 5 to 20 m (Frikken, 1999). The well-developed Westphalian fluvial sandstones of the Campine Basin, which continue underneath the Roer Valley Graben and the West Netherlands Basin, have so far been little explored for gas.

Surprisingly, fluvial channel sandstones from a poorly drained depositional setting tend to have a somewhat better reservoir quality than those from a well-drained setting. This may result from better sorting. In addition, channels in a poorly drained setting are closer to coal beds, and early gas charge from such coals may have played a role in porosity preservation.

Silesian sandstones deposited in a shallow-marine or lacustrine environment tend to be fine-grained with very

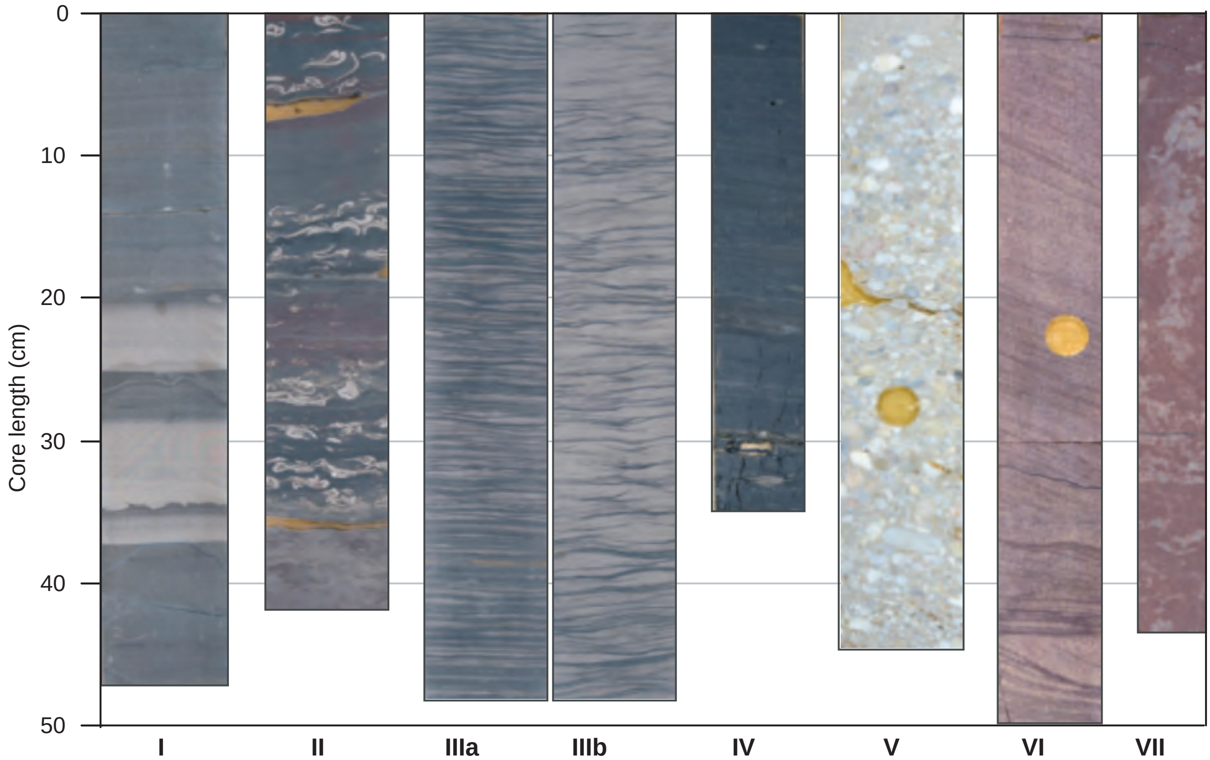

Fig. 10. Silesian of the Netherlands: facies variation as seen in cores. I) Marine shales with turbidites. Namurian-Westphalian A, well Blija-101. II) Marine shales with Brachiopods. Westphalian A, well Midwolda-1. III) Shallow marine: prograding mouth-bar deposits with flaser bedding (a) and linsen bedding (b). Westphalian A, well Stedum-1. IV) Coal. Westphalian A, well Zuidbroek-1. V) Braided-river conglomerates. Westphalian B/C, well D15-3. VI) Fluvial channel sandstones with trough cross-bedding. Westphalian D, well K2-1. VII) Mature paleosol (vertisol). Westphalian D, well Emmercompascuum-1. With thanks to Greg van de Bilt (PanTerra Geoconsultants).

low porosity. So far, in the Dutch offshore, no commercial finds have been made in such sandstones. However, the Namurian is still poorly known and could locally contain better reservoirs, as suggested by wells in the UK and outcrops in the UK and Germany.

A second important factor in the prediction of porosity in Silesian reservoirs is the maximum depth of paleoburial. In most of the Netherlands, this corresponds to the present-day depth. However, in some inverted basins, the Silesian has experienced net uplift since the Late Cretaceous, and its reservoir quality can be even poorer than its current depth would suggest. Examples are the Broad

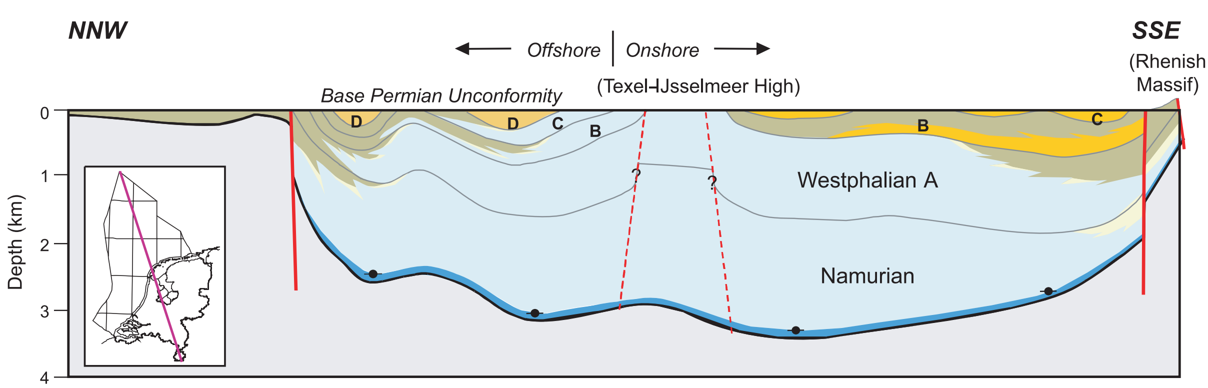

Fig. 11. Facies distribution: cross-section through the Silesian of the Netherlands (same as Fig. 1), flattened on the Base Permian Unconformity. Legend as in Fig. 8.

Fourteens Basin, the Central and West Netherlands basins and the southern margin of the basin, close to the London-Brabant Massif and the Rhenohercynian Zone.

Understanding the diagenesis is also crucial in the prediction of Silesian reservoirs. In general, diagenesis has had a polarising effect: layers with slightly higher primary porosity often also show a relatively higher secondary porosity. Finer-grained, less porous beds, especially when containing detrital clay, were at a disadvantage from the start, and often have ended up with completely occluded pores.

Most petrographic studies of Silesian sandstones show a complex diagenetic sequence involving cementation by quartz and sometimes carbonates, dissolution of feldspar and, most importantly, authigenic growth of kaolinite and illite. The authigenic clay is predominant where detrital clay was present in the first place. Such reservoirs may have a reasonable porosity, but a comparatively low permeability. Surprisingly, a positive correlation was observed between quartz cement and porosity. Quartz cementation tends to take place in the cleaner sandstones where less authigenic clay is present. The quartz cement often forms a coating around the grains. Thus it serves as fabric support and prevents compaction and grain dissolution of the reservoir rock during burial.

The depth below the Base Permian Unconformity does not stand out as an important factor in porosity prediction. Porosity enhancement may have occurred by leaching of Silesian rocks during the pre-Base Permian erosion. However, due to diagenetic overprinting by other events in the subsequent 250 Ma of burial, not much evidence of leaching below the unconformity was found.

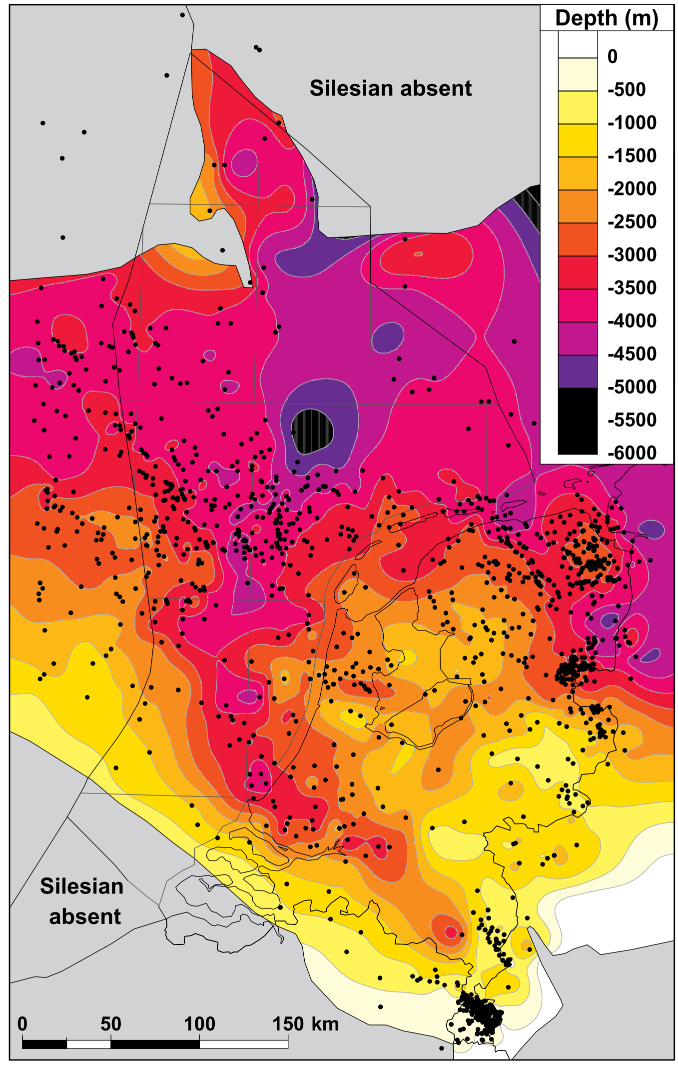

Fig. 12. Depth map of top Silesian, contoured from 1626 wells.

Tectonic development

Permian subcrop map

The subcrop map below the Base Permian Unconformity displays a complex erosional pattern, which resulted from the Late Carboniferous, Asturian inversion phase and Early Permian thermal uplift (Fig. 13). Toward the end of the Carboniferous, the total Silesian sediment column measured up to 5500 m. The inversion and associated truncation of the Silesian subsystem are most severe around the central E-W trending anticline in the northern Dutch onshore and the adjacent offshore area to the west. Here, up to 1000 m of Westphalian A, B and C and possibly 800 m of Westphalian D have been eroded. It is not clear whether Stephanian was deposited in this area and subsequently eroded. At the northern fringe of the Carboniferous Basin, on the Mid North Sea High, Dinantian and older sediments subcrop below the Base Permian Unconformity. Seismic stratigraphy and fission-track analysis suggest that some 1250 m of Silesian were originally present in this area.

Structurally, there is a close correspondence between high or low areas at Base Permian and later tectonic features (Geluk, this volume; De Jager, this volume). The high areas, which were affected by deep erosion of Carboniferous strata during the Late Carboniferous and Early Permian, often reappear as horsts in later times. The low areas, less affected by erosion, tend to reappear as grabens. Examples of such reactivation in the Dutch onshore are the West Netherlands Basin, the Lauwerszee Trough, the western part of the Lower Saxony Basin and the Peel Block. Examples from the offshore include the Elbow Spit High and the Step Graben.

During the Middle and Late Permian, after the Early to Middle Permian uplift and erosion, Rotliegend sediments covered most of the Carboniferous in the Netherlands. The Base Permian Unconformity represents in most places a time gap of some 40 to 60 Ma (Geluk, this volume).

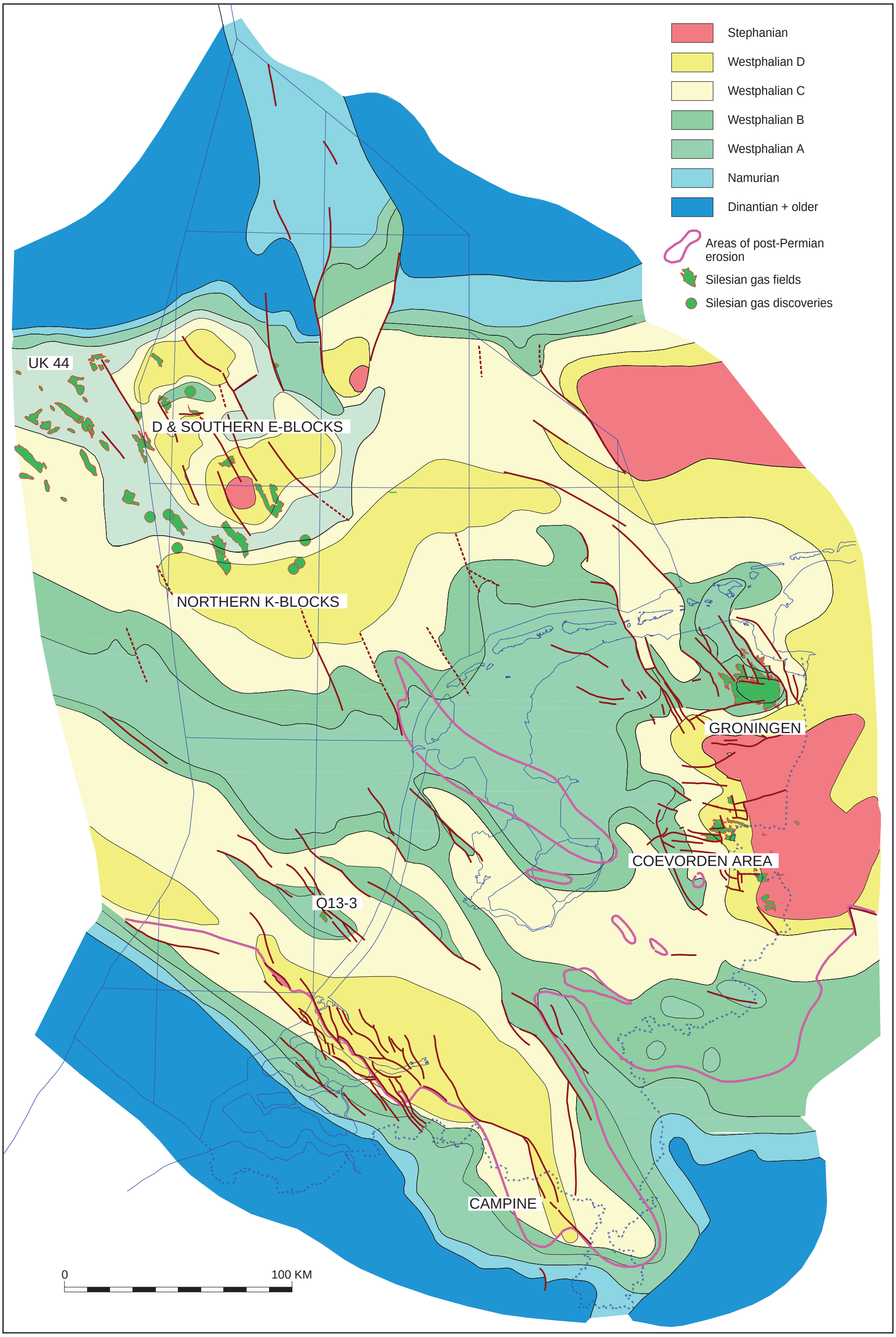

Fig. 13. Subcrop map of the Base Permian Unconformity, also showing gas accumulations in Silesian reservoirs. Where the Permian has been eroded, or not deposited, the configuration of top Carboniferous under younger unconformities is shown.

Silesian unconformities and burial history

A clear onlap of seismic reflections in the Namurian A can be observed against both the London-Brabant Massif and the Elbow Spit High. Updip of these onlapping reflections, the evidence for a Lower Namurian Unconformity has generally been removed by Early Permian and Mesozoic truncation.

Regional seismic data in the Netherlands show little evidence for intra-Silesian unconformities. Most seismic lines display a parallel, 'tram-line' pattern within the Silesian, often with a marked truncation at the Base Permian Unconformity (Fig. 5). Detailed isopach maps and well data occasionally suggest some degree of synsedimentary tectonics. However, the general rate of subsidence was so high and steady that there was little time to form unconformities. Only at those margins of the basin, where subsidence was relatively low, erosional unconformities may have formed as a result of a base-level fall. Incised fluvial valleys, at (para-)sequence boundaries, can be expected in this setting. An example of the latter possibly occurs in offshore well B17-4, where Westphalian B directly overlies Namurian. In UK quadrant 53, in the southern part of the basin, actual truncation of Westphalian B-C sediments below Westphalian C-D has been observed on seismic data and in wells (Tubb et al., 1986). The event is explained by a heat-flow phenomenon and associated basin-edge tectonics and is known as the 'Symon Unconformity'. Another time-equivalent unconformity has been observed in the northern part of the basin, in UK quadrants 43 and 44, where coarse-grained Westphalian C-D sediments onlap against Westphalian B-C sediments (Besly, 1998).

In the German and the British offshore, especially in the vicinity of the Mid North Sea High, a Base Stephanian Unconformity is frequently reported (e.g. Ziegler, 1990). This unconformity may also be present in the Dutch northern offshore. However, in the absence of borehole penetrations with unambiguous biostratigraphic dating and of truncating seismic reflections, no hard evidence can be presented to support this.

The Base Permian Unconformity is generally the deepest that can be recognised in the Dutch on- and offshore. This is related to regional uplift of mainly Permian age. Isochore correlation suggests that in much of the Netherlands up to 1800 m of Silesian have been eroded during the Early to Middle Permian (Fig. 1). From the Middle or Late Permian onward, younger sediments steadily buried Silesian rocks. Uplift, with sometimes renewed erosion of Silesian strata, only took place in limited areas during the Late Jurassic to Early Cretaceous, Kimmerian rifting and during the Late Cretaceous to Early Tertiary, Subhercynian and Laramide inversion phases. Especially on the Texel-IJsselmeer High and in the southern Netherlands, the Silesian and its overburden were extensively eroded. Coalification studies indicate that in southern Limburg, where the Westphalian is close to the surface, up to 5400 m of Silesian may have been removed (Veld et al., 1996).

Igneous rocks

Igneous rocks are fairly common in the Silesian of the Netherlands. The first well in the country to recover oil, Corle-1 (eastern Gelderland, 1923), penetrated a dolerite intrusive within the Westphalian A. Thiadens (1963) described seven further occurrences of igneous rock within the Silesian, and many more have been found since (Sissingh, 2004; Van Bergen & Sissingh, this volume).

Understanding and information about these igneous rocks are important for three reasons: i) they help to unravel the tectonic history, ii) extrusive layers form important correlation horizons and can be used for numerical dating, and iii) knowing why and predicting where igneous rocks may occur can avoid unpleasant exploration surprises: igneous rocks, like hydrocarbon occurrences, often show similar strong seismic-amplitude anomalies.

Intrusive rocks

The compilations of Sissingh (2004) and Van Bergen & Sissingh (this volume) show that intrusive rocks are rather common in the Silesian in the east of the country and in the northern offshore area. Radiometric data indicate Carboniferous, Early to Middle Permian, Late Triassic and Late Jurassic ages. Intrusions for which these data indicate a possible Silesian age are present in wells Dwingelo-2 (Drenthe) and Nagele-1 (Flevoland). These intrusions were probably emplaced at rather shallow depth, relatively soon after sediment deposition (RGD, 1993). The intrusion in Dwingelo can be followed on seismic data over several kilometres.

The most prominent example of a younger intrusive within Silesian sediments is the 72-m-thick dolerite in well Winterswijk-1 (eastern Gelderland; 4077–4149 m along hole). This dolerite has a K/Ar dating of 218 ± 6 Ma (i.e. Late Triassic) and intruded the Lower Namurian. It caused a strong seismic-amplitude anomaly, and was mistakenly interpreted as Top Dinantian when the well was proposed.

Extrusive rocks

Only a single occurrence of unequivocally extrusive rock is known from the Carboniferous of the Netherlands. In the cored interval of well Steenwijkerwold-1 (Overijssel), an 8-m-thick basaltic layer is present, with soft-sediment deformation in the underlying sediments, and evidence of weathering and subsequent submergence at its top (Fig. 14). The overlying sediments contain a well-preserved palynological assemblage, indicating a Late

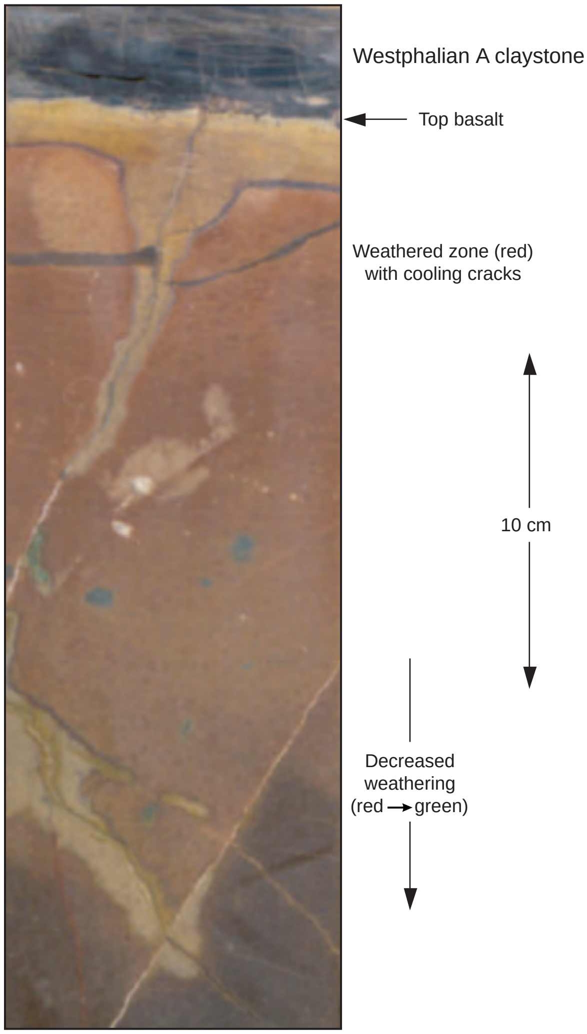

Fig. 14. Core photograph: top of Steenwijkerwold lava flow. Undisturbed lacustrine claystone with well-preserved palynoflora indicating a Late Westphalian A age, overlying an 8-m-thick weathered basalt layer. Well Steenwijkerwold-1, 1936 m along hole.

Westphalian A age. The shallow intrusives in Dwingelo-2 and Nagele-1 are probably related to the same volcanic phase, which may be linked to Variscan transtension along the proto-Texel-IJsselmeer High. This would indicate that the boundary faults of this high pre-date by far the Kimmerian tectonic phase in which they were most active. The 'Kaolin-Kohlentonsteine' described from the Westphalian A-C of the German Ruhr area (e.g. Burger, 1982) and the southern Netherlands (Kimpe, 1962) may be correlated with the same volcanic phase. Another example from the Netherlands is a 5-cm-thick ash layer cored at 3134 m along hole in well Hoogenweg-1 (Overijssel). A study by the Rijks Geologische Dienst (RGD, 1988) attributed this layer to the 'Hagen-1 Tonstein' of Early Westphalian C age, as defined by Burger (1982). Active volcanism during the Westphalian A-C has also been described from the UK, for example by Kirton (1984).

Further Silesian extrusives may occur offshore in the area of the Step Graben and Elbow Spit High. The evidence for these is indirect only and consists of seismic reflection data and the presence of Namurian extrusives in nearby German offshore wells.

Economic geology

Coal and gas, volumetrics

The Silesian is the source for almost all coal and gas extracted in the Netherlands, and as such is of enormous economic importance. Based on the isopach maps in Figure 9, the total volume of Silesian deposits in the Dutch on- and offshore is estimated at 231 000 km^3. This corresponds to an average thickness of 2.35 km. Average coal percentages per stage are estimated between 0.1 and 2.1%, based on in-house studies of the Nederlandse Aardolie Maatschappij. These percentages can be considered conservative. Other authors have given higher estimates of up to 3.5% (e.g. Van Wijhe & Bless, 1974; Hedemann et al., 1984). However, these estimates have been based on onshore observations along the southern basin margin, whereas our facies maps (Fig. 8) suggest lower coal percentages for the offshore. The calculated volumes of rock, coal and combustible gas (Table 1) have to be considered a crude approximation, but their order of magnitude is probably correct. The average density of coal is assumed to be 1.5 g/cm^3, i.e. low-volatile bituminous to anthracitic. It is assumed that 1 tonne of this type of coal has generated some 136 m^3 of combustible gas (Rightmire, 1984).

Resources

The Silesian is important for the Dutch economy in the following ways.

PAST

Coal Coal has been mined in the Kerkrade area since medieval times (Bless et al., 1984). Larger-scale commercial coal production in southern Limburg started in 1847 and ended in 1974. During those 128 years, an estimated 582 Mt of coal were produced in several mines, all in southern Limburg (Westen, 1971; Stuffken, 1987; Van Bergen et al., this volume). The calculations in Table 1 suggest that some 2.3×10^{12} tonnes of coal are present in the Dutch subsurface. Thus, ca. 0.025% of the Dutch coal has been mined. Most of the remainder is too deep to be accessible. At depths < 1 km, coal occurs only in southern Limburg, the Peel Block and the Achterhoek (east Gelderland). The latter two areas

Table 1. Estimated rock, coal and combustible-gas volumes for the Silesian in the Dutch on- and offshore.

Age	Rock volume (10^3 km^3)	Coal % (av. est.)	Coal volume (km^3)	Coal weight (10^9 tonnes)	Combustible gas generated (10^{12} m^3)
Westphalian D	3	0.1	3	5	1
Westphalian C	7	1.6	114	170	23
Westphalian B	23	2.1	477	715	97
Westphalian A	82	1.0	820	1230	167
Namurian	116	0.1	116	174	24
Total	**231**		**1530**	**2294**	**312**

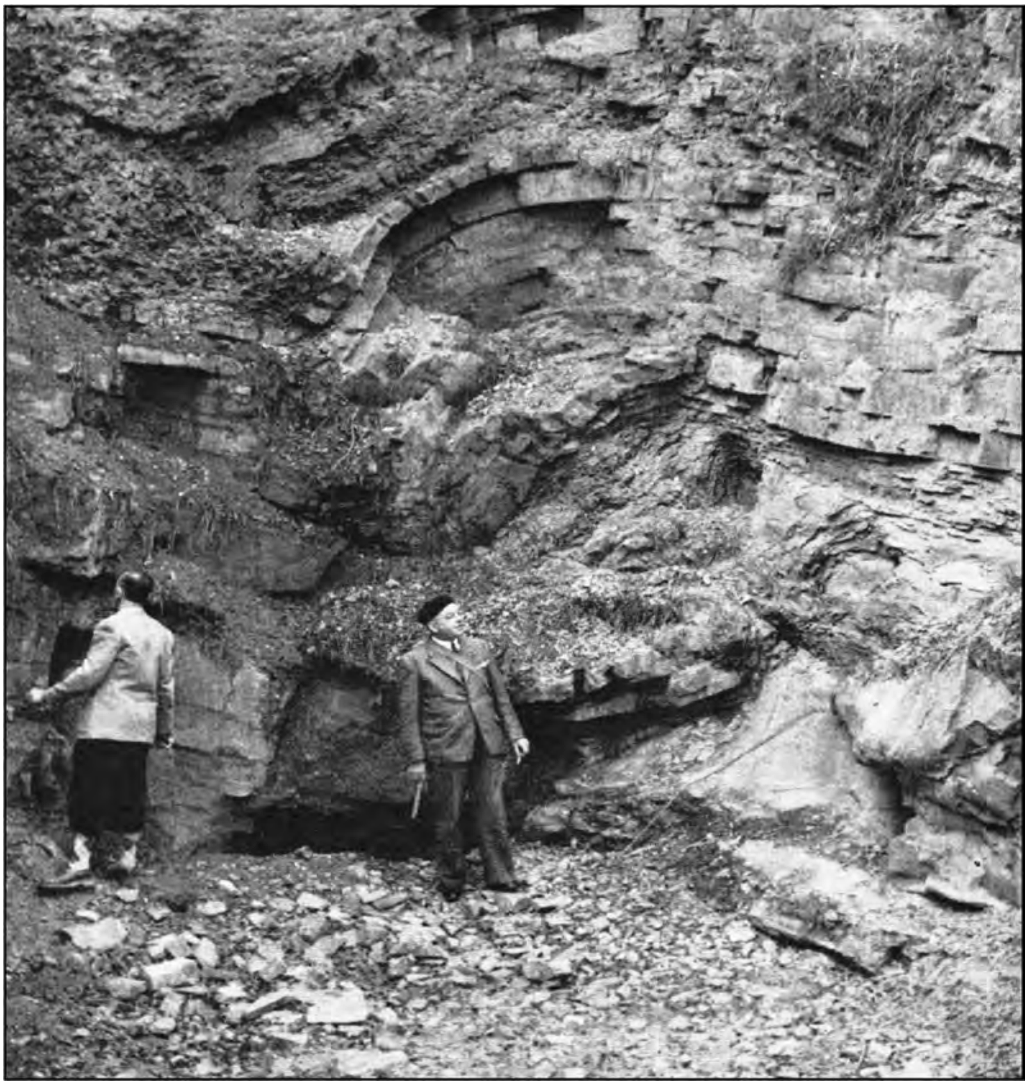

Fig. 15. Example of Silesian outcropping in the Netherlands: sandstone with coal streaks (probably Westphalian A, Baarlo Formation) in the Kamp Quarry, Geul valley (Limburg), around 1960. From Krul (1963). Reprinted with permission from Thieme-Meulenhoff Publishers, Zutphen.

have been subject to extensive coal-exploration campaigns by the Geological Survey, but the results never led to production (Peelcommissie, 1963; Van Tongeren, 1987; Van Bergen et al., this volume).

Building stone In the Meuse valley, between the cities of Maastricht and Roermond, Silesian sandstone used to be the principal building stone from Roman times up to the Romanesque period. Much of this was imported from

Germany and Belgium. In the Netherlands, a few small abandoned quarries can still be seen in the Geul valley in southern Limburg, where they serve as geological monuments (Fig. 15; Bosch, 1989). The occurrence of ganister, that may be used in furnaces, has been described from Dutch coal mines by Kimpe (1963). Ganister is a very hard non-metamorphic quartzite developed in the seat earth below coal seams.

PRESENT

Gas source Combustible gas is unquestionably the main Silesian resource in commercial terms. It is estimated that over 95% of all such gas discovered in the Netherlands was generated from Westphalian coal. The remainder originated from other source rocks, mainly the Jurassic Posidonia Shale (De Jager & Geluk, this volume). The calculations in Table 1 suggest that more than 300×10^{12} m^3 of combustible gas may have been generated in the Netherlands over geological time. Less than 1.5% of this volume has been discovered. It is safe to assume that an overwhelming majority (> 95%) of the generated gas has escaped through surface seeps, mostly in the distant past.

Oil source Most of the oil discovered in the Netherlands has been correlated with the Posidonia Shale, but oil sourced from the Silesian may have been found in the West Netherlands Basin in wells Papekop-1 and Ottoland-1 (De Jager et al., 1996; Van Balen et al., 2000).

Gas reservoir Over 140×10^9 m^3 of combustible gas have been discovered in Silesian reservoirs in the Netherlands during the last 55 years. Some 65% of this volume was found in the Coevorden area, which also produces from the Zechstein, and in the Groningen field, where the gas column extends from the Rotliegend down into the Westphalian. The remainder was discovered offshore in fields in the southern D and E, and northern K blocks (Fig. 13). The latter also produce from the Rotliegend.

FUTURE

Coalbed methane In southern Limburg, the Peel Block and the Achterhoek, commercial extraction of coalbed methane may be possible in the future. At present, however, such extraction cannot yet compete with conventional gas production. An extended test just across the border in Belgium (well Peer-1, 1992) had problems with water production and no economic opportunity was identified. The deep burial and subsequent structural inversion in the Campine Basin appear to have resulted in coals that are undersaturated in methane (Van Bergen et al., this volume).

The power plant in Buggenum (Limburg) operates on gasification of imported coal aboveground.

CO_2 sequestration Recent investigations by TNO indicate that CO_2 storage in Westphalian coal could be an option for the Netherlands to fulfil part of its obligations under the Kyoto protocol of 1997 to reduce net CO_2 output. The investigations also address the possibility of stimulating coalbed-methane production by CO_2 sequestration (Van Bergen et al., this volume).

ACKNOWLEDGEMENTS

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Permian

M.C. Geluk

ABSTRACT

Permian deposits in the Netherlands are represented by the Lower Rotliegend, Upper Rotliegend and Zechstein groups. The Lower Rotliegend Group (Middle Permian), of volcanic origin, is present only locally. The Upper Rotliegend and Zechstein groups, Middle to Late Permian in age and deposited under warm and arid climatic conditions, are present throughout most of the Netherlands. The Upper Rotliegend Group represents fluvial, eolian and playa-lake deposits, with the lake situated in the northern offshore area. The Zechstein Group comprises a series of marine evaporites and carbonates, which show a gradual retreat of the sea during its deposition. By the end of this deposition, continental and more humid conditions had returned. The depositional thickness of the Permian reaches almost 2000 m in the northern offshore. The regional unconformity at the base of the Permian represents in most places a hiatus of 40 to 60 Ma. The Permian deposits contain more than 95% of the large reserves of natural gas in the Netherlands; they also contain exploitable rock salt and potassium-magnesium salts. The thick layer of Zechstein rock salt deformed into a large number of salt pillows and diapirs, and strongly influenced the post-Permian structural development of the country.

Keywords: Netherlands, Rotliegend, Zechstein, desert, volcanics, evaporites, salt, natural gas

Introduction

The Permian was deposited after the Late Carboniferous final stage of the Variscan orogeny. Permian rocks contain important hydrocarbon reservoirs and seals as well as exploitable salt deposits. In the Netherlands and the adjacent southern North Sea the Permian rests unconformably upon relatively mildly deformed Namurian to Stephanian sedimentary rocks. South-east of the Netherlands, in the area of the Variscan Mountains, Permian strata rest upon strongly deformed Paleozoic sediments (Ziegler, 1990).

In the Netherlands, a hiatus represents the entire Early Permian. In Middle and Late Permian times, the Netherlands became included in a large E-W trending complex of sedimentary basins, usually referred to as the Southern Permian Basin, stretching from the UK into Poland (Figs 1, 2; Ziegler, 1990; Glennie, 1997, 1998; Lokhorst, 1998; Geluk, 2005). The sedimentation area gradually expanded during the Mid and Late Permian until the London-Brabant Massif and the Rhenish Massif formed its southern, and the Mid North Sea High and Ringkøbing-Fyn High its northern margin. The Ems Low

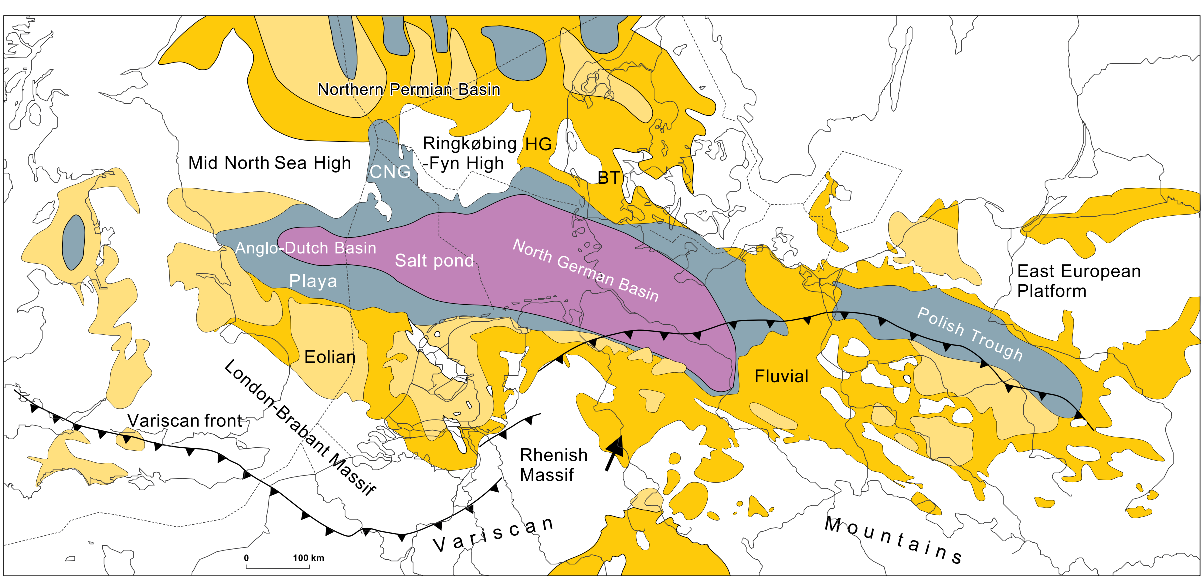


Fig. 1. Present-day distribution and facies map of the Upper Rotliegend Group (late Middle to early Late Permian) in the Southern Permian Basin (Geluk, 2005). BT: Bramble Trough; CNG: Central North Sea Graben; HG: Horn Graben.

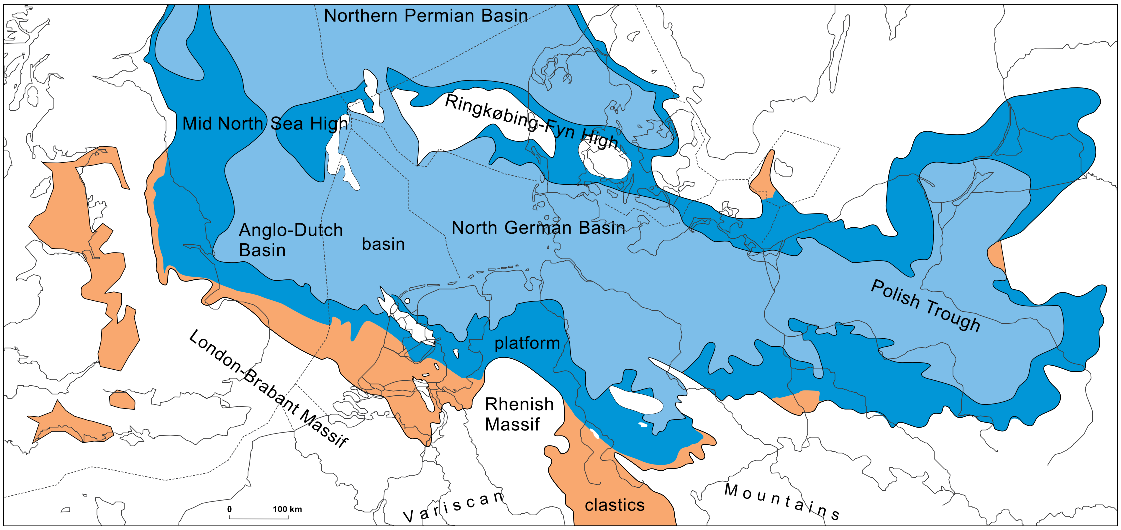


Fig. 2. Present-day distribution and facies map of the Zechstein Group (Z2 Carbonate, Late Permian) in the Southern Permian Basin (after Lokhorst, 1998; Taylor, 1998; Geluk, 2005).

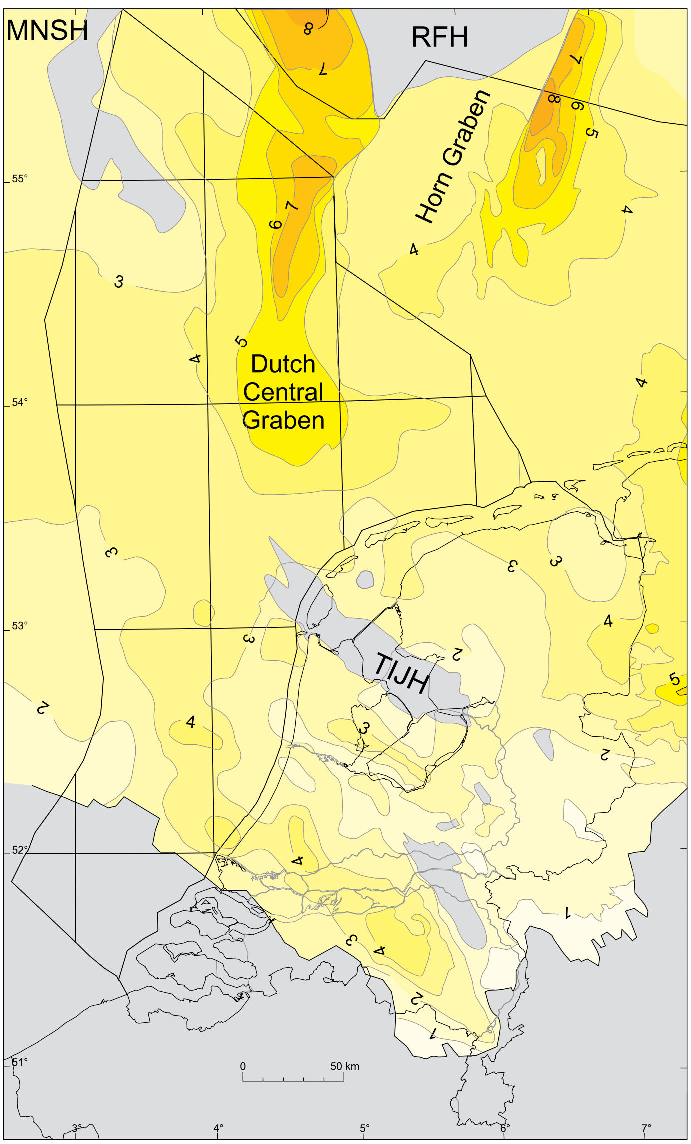


Fig. 3. Depth map of the base of the Zechstein Group (in kilometres). Compiled from the geological atlas of the Netherlands (unpublished NITG compilation), with additional data from Day et al. (1981), Bailey et al. (1993), Vejbaek & Britze (1994), Kockel (1995) and Baldschuhn et al. (1996). TIJH: Texel-IJsselmeer High; MNSH: Mid North Sea High; RFH: Ringkøbing-Fyn High.

and the Dutch Central Graben formed areas of volcanic activity during the Middle Permian, and topographic lows in Late Permian times. During the Middle Permian the differential subsidence of the future Broad Fourteens and Central Netherlands basins started. Sediment supply originated mainly from the Variscan Mountains in the south and to a minor extent also from the highs north of the basin.

The depositional thickness of the Permian increases northwards from less than 50 m in the southern Netherlands to almost 2000 m in the northern offshore area. Post-Permian erosion and salt movement strongly influenced the original thickness patterns. A depth map of the base of the Permian based on well data is shown by Van Buggenum & Den Hartog Jager (this volume). The depth of the base of the Zechstein Group ranges from less than 1 km in the eastern Netherlands to over 7 km in the Dutch Central Graben, and over 8 km in the Danish Central Graben and the Horn Graben (Fig. 3).

Stratigraphy

The Permian in the Netherlands is divided into three groups (Fig. 4; Van Adrichem Boogaert & Kouwe, 1994):

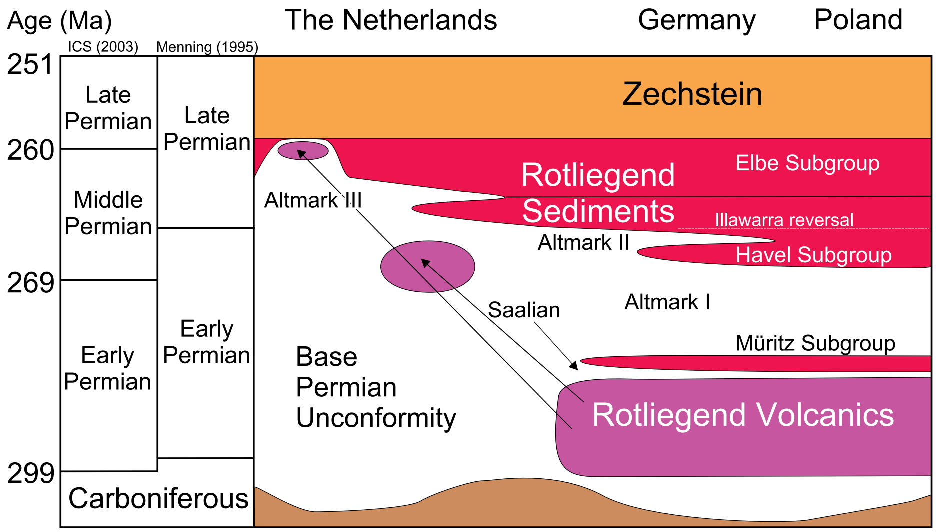


Fig. 4. Summary stratigraphy of the Permian in the Netherlands and adjacent areas (after Lokhorst, 1998; Glennie, 1998; Geluk, 1999), including a correlation with the ICS (2003) subdivision. Saalian and Altmark I, II and III refer to unconformities in the succession. In most of the Netherlands a single mega-unconformity, the Base Permian Unconformity, separates the Middle to Upper Permian from the Carboniferous. Note the occurrence of two distinct volcanic pulses in time.

1. the Lower Rotliegend Group, a succession of volcanic and clastic rocks with a geographically limited distribution;
2. the Upper Rotliegend Group, comprising fine-grained clastics and evaporites in the northern offshore and predominantly sandstones in the onshore and western offshore area;
3. the Zechstein Group, comprising marine evaporites, carbonates and subordinate clastics.

The contact between the Lower Rotliegend Group and the overlying Upper Rotliegend or Zechstein groups is unconformable, whereas the contact between the Upper Rotliegend and the Zechstein Group is conformable.

Traditionally the Permian deposits were subdivided into a Lower Permian Rotliegend and an Upper Permian Zechstein. Until recently, the Rotliegend was considered to represent the Early Permian, and the Zechstein the Late Permian (Menning, 1995; Plein, 1995; Glennie, 1998). Radiometric dating of the Lower Rotliegend Group in Germany documents an Early Permian age (Plein, 1995; Stemmerik et al., 2000), but Middle Permian ages in the Netherlands (Sissingh, 2004; Geluk, 2005). The Upper Rotliegend Group is almost completely devoid of fossils, having been deposited under hypersaline or oxidizing conditions. Here, non-biostratigraphic tools such as chemostratigraphy, magnetostratigraphy and sequence stratigraphy are applied (George & Berry, 1994; Yang & Baumfalk, 1994; Schuurman, 1998). The Zechstein Group, especially the carbonates and claystones, yields palynomorphs and marine fossils (bryozoa, bivalves, algae).

Menning (1995) published a new view on the age of the Rotliegend using paleomagnetic data. The magnetic Illawarra reversal, observed within the Upper Rotliegend in Germany, almost coincides with the boundary between the Early and Late Permian. In Germany, a thick, more complete Upper Rotliegend succession is present comprising both Early and Late Permian deposits, whereas in the Netherlands a much thinner succession comprises only Late Permian deposits (Fig. 4). The Early Permian part of the Upper Rotliegend is often referred to as the Upper Rotliegend I, the Late Permian part as the Upper Rotliegend II. Various authors have adopted Menning's view for the Netherlands and the southern North Sea (Van Adrichem Boogaert & Kouwe, 1994; Plein, 1995; Glennie, 1997, 1998; Geluk, 1999, 2005). A shorter duration of the time of deposition for the Upper Rotliegend in the Netherlands than previously assumed had also been deduced by Yang & Baumfalk (1994) on the basis of high-resolution sequence stratigraphy.

The Permian Stratigraphic Committee agreed upon a three-fold subdivision of the Permian (IUGS, 2000; ICS, 2003). Following this subdivision, the Lower Rotliegend Group is Early to Mid Permian in age, and the hiatus between the Upper Rotliegend and the Carboniferous in the Netherlands spans the entire Early and part of the Middle Permian. The Upper Rotliegend is Middle to early Late Permian, the Zechstein Group Late Permian.

A different approach to subdivision of the deposits is sequence stratigraphy, where time-lines are correlated. This results in a better understanding of the depositional relations of the sediments. Sequence-stratigraphic models of the Upper Rotliegend Group have been published by George & Berry (1994), Yang & Baumfalk (1994) and Plein

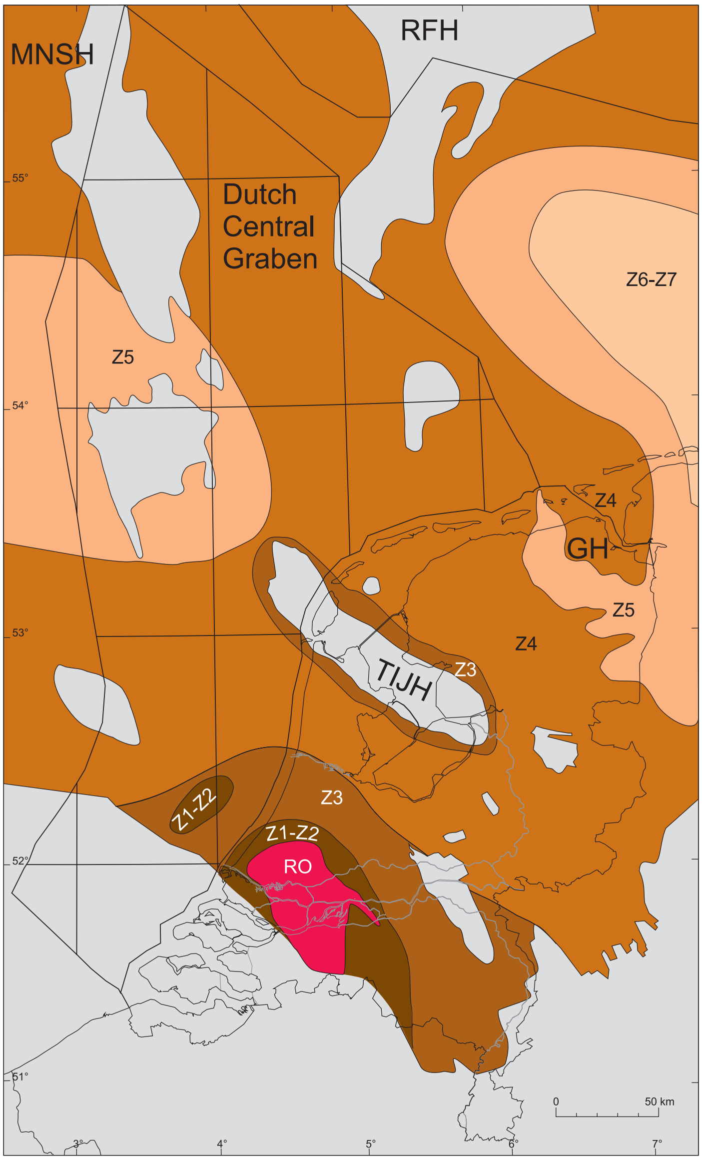


Fig. 5. Subcrop map below the Zechstein Upper Claystone Formation (after Geluk et al., 1997; Best, 1989). The present-day distribution of this formation is shown; it is absent in the grey areas. In the area of the future West Netherlands Basin the Upper Rotliegend Group (RO) subcrops below the Zechstein Upper Claystone Fm. Z1: Z1 (Werra) Fm, Z2: Z2 (Stassfurt) Fm, Z3: Z3 (Leine) Fm, Z4: Z4 (Aller) Fm, Z5: Z5 (Ohre) Fm. Z6-Z7 represent the youngest evaporite cycles of the Zechstein Group identified in Germany (Best, 1989). GH: Groningen High; other abbreviations as in Fig. 3.

(1995); all models propose wet-dry Milankovitch-driven climatic cycles, with a periodicity of 100 and 200 ka, to control the sedimentation. These high-resolution sequences are grouped in larger-scale cycles, with a duration of between 0.7 to 1.4 Ma. The cyclicity of the Zechstein has been described traditionally as governed by initial transgressions, followed by regressive phases with evaporite deposition (Richter-Bernburg, 1955). These cycles, bounded by maximum flooding surfaces, are called genetic stratigraphic sequences (Galloway, 1989). A more modern view was published by Tucker (1991), who proposed an evaporite-carbonate sequence stratigraphy with sequence boundaries at the tops of the carbonates. The onset of the Zechstein type of cyclicity lies already in the upper part of the Rotliegend.

Regional correlations

The Permian in the Netherlands can be correlated without major problems to the adjacent countries (Geluk, 2005). The equivalents of the volcanoclastics of the Lower Rotliegend Group are known as the Karl Formation in Denmark and the adjacent UK central North Sea (Stemmerik et al., 2000), and as the Altmark Group in Germany (Plein, 1995). Radiometric ages of these groups, however, do vary (see below).

The Upper Rotliegend Group correlates with the Auk Formation in the UK central North Sea, and the Silverpit and Leman formations of the UK southern North Sea (Johnson et al., 1994). The equivalent in Germany is the Hannover Subgroup (Upper Rotliegend II of Plein, 1995; Verdier, 1996); equivalents of the underlying formations do occur only in Germany and Poland (Geluk, 2005). Equivalents of the Upper Rotliegend Group have been described as the so-called Zechstein conglomerate in north-east Belgium (De Craen & Swennen, 1992) and in the Lower Rhine coal-mining area in Germany (Hilden, 1988). These are considered time equivalents of only the youngest Upper Rotliegend or even of the oldest Zechstein sediments.

The Zechstein Group in the Netherlands correlates with the Zechstein Supergroup of the UK southern North Sea and with the German Zechstein. Most of the unit names are identical. A major discrepancy, however, affects the definition of the top of the group between the various countries. This top is picked considerably lower in the UK than in the Netherlands and Germany; as a result, rocks considered part of the Permian Zechstein in the latter two countries are assigned to the Triassic Bacton Group in the UK southern North Sea (Geluk et al., 1996; Lokhorst, 1998; Fisher & Mudge, 1998; Taylor, 1998; Geluk, 1999, 2005).

Tectonic setting and events

The deposition of the Permian followed the Variscan orogeny. The compressive movements of this orogeny ended during the Westphalian D (Ziegler, 1990). The deformation of the Carboniferous to the north of the Variscan Front in the Netherlands post-dates the Variscan folding and has been attributed to Early Permian wrenching (Van Wijhe et al., 1980; Van Wijhe, 1987; Ziegler, 1990). It resulted in a major unconformity between the Upper Rotliegend and older deposits. This unconformity is referred to as the Base Permian Unconformity. In north-

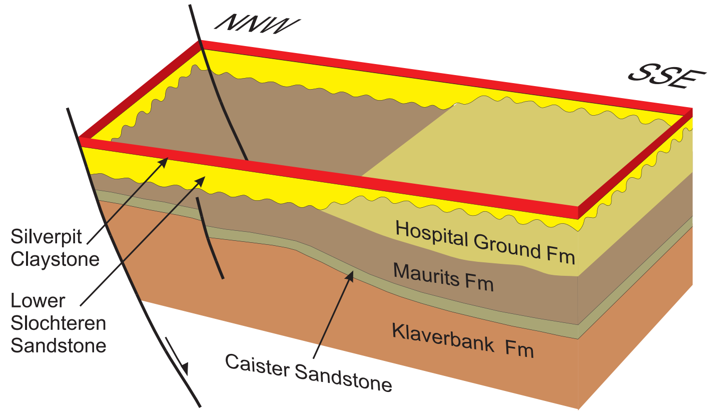


Fig. 6. Model of the relationship between the subcrop at the Base Permian Unconformity and the Upper Rotliegend sandstone distribution in the north-west offshore area (Geluk & Mijnlieff, 2001; Geluk et al., 2002). Sand-prone Carboniferous units (Hospital Ground Fm., Caister Sandstone) formed topographic ridges on which only thin Lower Slochteren sandstones were deposited. Fine-grained coal-bearing Carboniferous successions (Maurits Fm.) formed topographic lows, where thick Lower Slochteren sandstones accumulated. The relief was up to 25 m.

ern Germany and Poland this wrenching was accompanied by widespread in- and extrusive volcanism along deep-seated fault zones. It affected the Netherlands only marginally.

Several minor tectonic pulses characterize the Late Permian (Gast, 1988; Plein, 1995; Glennie, 1998; Geluk, 1999). These are thought to herald the Permian breaking-up of Pangea (Vai, 2003). The pulses occurred prior to and during the deposition of the Upper Rotliegend Group (the Saalian and Altmark I to III pulses; Plein, 1995) and during deposition of the Zechstein (Tubantian I; Geluk, 1999). As a result, a connection was established between the Southern Permian Basin and the Barents Sea. The Saalian and Altmark pulses influenced the distribution of basal Upper Rotliegend sands (Van der Baan, 1990; Van de Sande et al., 1996; NITG, 1998). The Tubantian II pulse is thought to be of compressional origin and related to the Uralian orogeny (Geluk, 2005), and marks the second, final, consolidation of Pangea at the end of the Permian (Vai, 2003).

There is evidence for considerable differential fault movement during the deposition of the Z1 (Werra) Formation (Wolf, 1985; Ziegler, 1989; Geluk, 1999): the Tubantian I phase. The faulting concentrated on the anhydrite platforms at the margins of the Zechstein basin. Loading of rapidly deposited anhydrite on a differentiated substrate is thought to have amplified these movements (Geluk, 1999). A series of halfgrabens and pull-apart-type basins formed, with fault offsets of up to 250 m in the Central Netherlands Basin. During this faulting the first fault–dip-closed traps of Upper Rotliegend sandstones were created, only 3 to 5 Ma after their deposition.

The Tubantian II movements during deposition of the upper Zechstein are more difficult to identify. These movements led to an unconformable contact between the Zechstein Upper Claystone and older Zechstein rocks (Fig. 5). They resulted mainly in tilting and uplift, and have

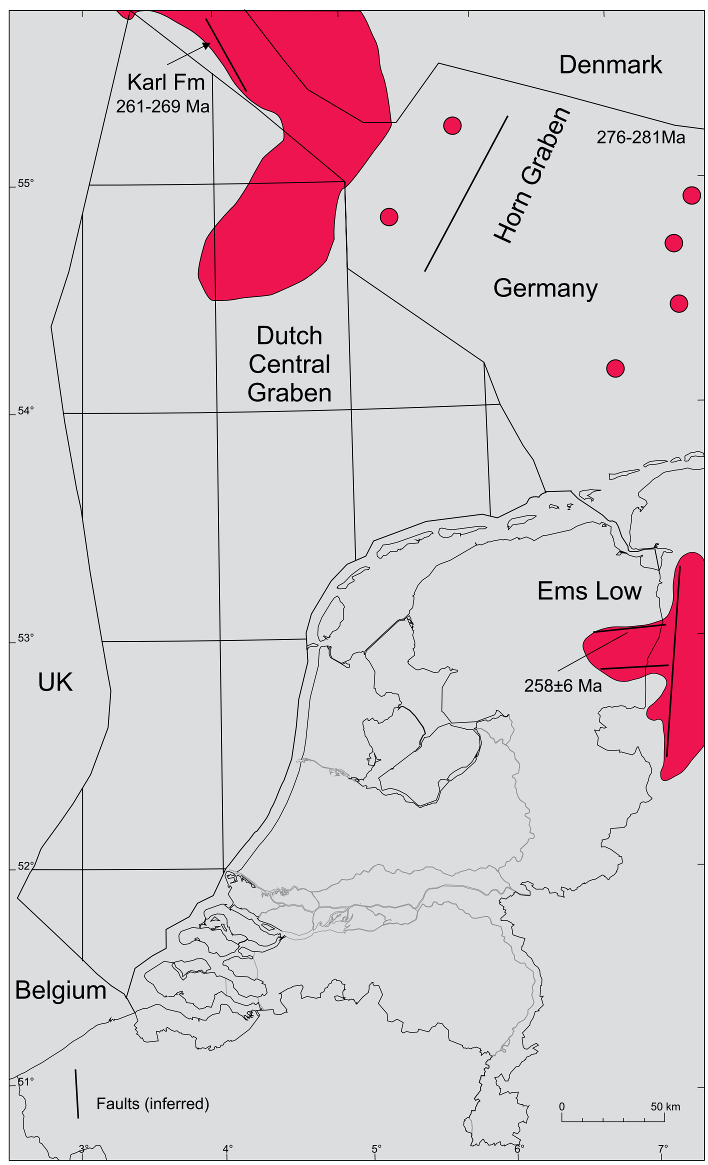


Fig. 7. Distribution of the Lower Rotliegend Group (after Lokhorst, 1998). Volcanic activity occurred in the Ems Low and the Dutch Central Graben. Red dots indicate isolated occurrences of volcanics in wells in the German sector of the North Sea. The eastern limit of the group in the Dutch Central Graben is not well-established. From the distribution of the volcanics a bimodal pattern for deep-seated faults can be inferred.

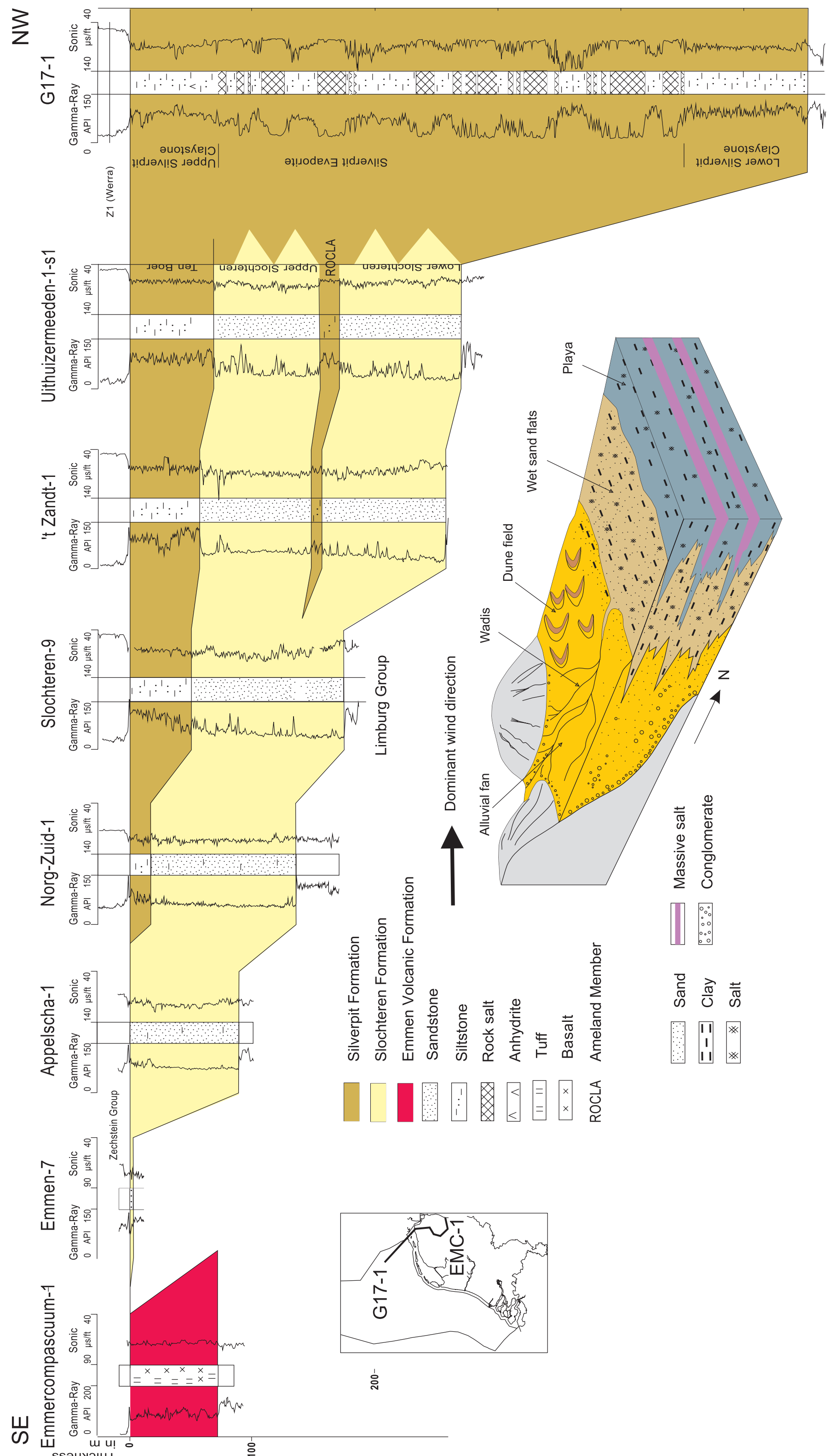


Fig. 8. Log correlation of the Lower and Upper Rotliegend groups, showing the transition from the sandy Slochteren Formation in the onshore area of the Netherlands into the playa of the Silverpit Formation in the northern offshore (G17-1). The reference level is the base of the Zechstein Group. A block diagram illustrates the depositional facies of the Upper Rotliegend Group, roughly along a similar transect. ROCLA: Ameland Member.

been interpreted by Geluk (2005) as being related to mild compression. Source areas south of the basin were uplifted, and sands were shed into the basin in the western offshore area and the UK sector (Z4 or Hewett Sandstone). In the main basin a series of halites were laid down (Z6-Z7; Best, 1989), which possibly are in part redeposited Zechstein salts.

Base Permian Unconformity

The hiatus comprising the entire Early Permian is one of the most important unconformities in the regional petroleum geology of the southern North Sea (De Jager & Geluk, this volume). Where the Upper Rotliegend overlies Namurian to Stephanian deposits, as is the case in most of the Netherlands, this hiatus represents 40 to 60 Ma. It should be taken into account that the unconformity is an amalgamation of several unconformities into a single mega-unconformity (Glennie, 1998). The name Saalian Unconformity, often used (Van Wijhe, 1987), is not correct since it refers only to one of the unconformities involved. The name Base Permian Unconformity is proposed for this mega-unconformity (Geluk, 2005; Fig. 4).

Despite the large gap in sedimentation, the unconformity may be very difficult to identify in some areas, because of a similar red-bed character across this boundary and the general absence of fossils. A combination of various methods, e.g. sequence stratigraphy, core studies and chemostratigraphy, may result in a proper identification (Schuurman, 1998). The paleotopography of this unconformity affected the sediment dispersal of the basal Upper Rotliegend sandstones. The relief of this topography developed in response to resistance to weathering of the subcropping units (sand-prone Carboniferous deposits forming ridges and coal-bearing fine-grained units forming lows) and in areas close to the Variscan Front in relation to faults (Geluk & Mijnlieff, 2001; Geluk et al., 2002; Fig. 6). In areas with sufficient sand thickness in the Upper Rotliegend the effects of this topography will hardly be noticed, but they are considered critical in areas near the limit of Upper Rotliegend sandstone distribution.

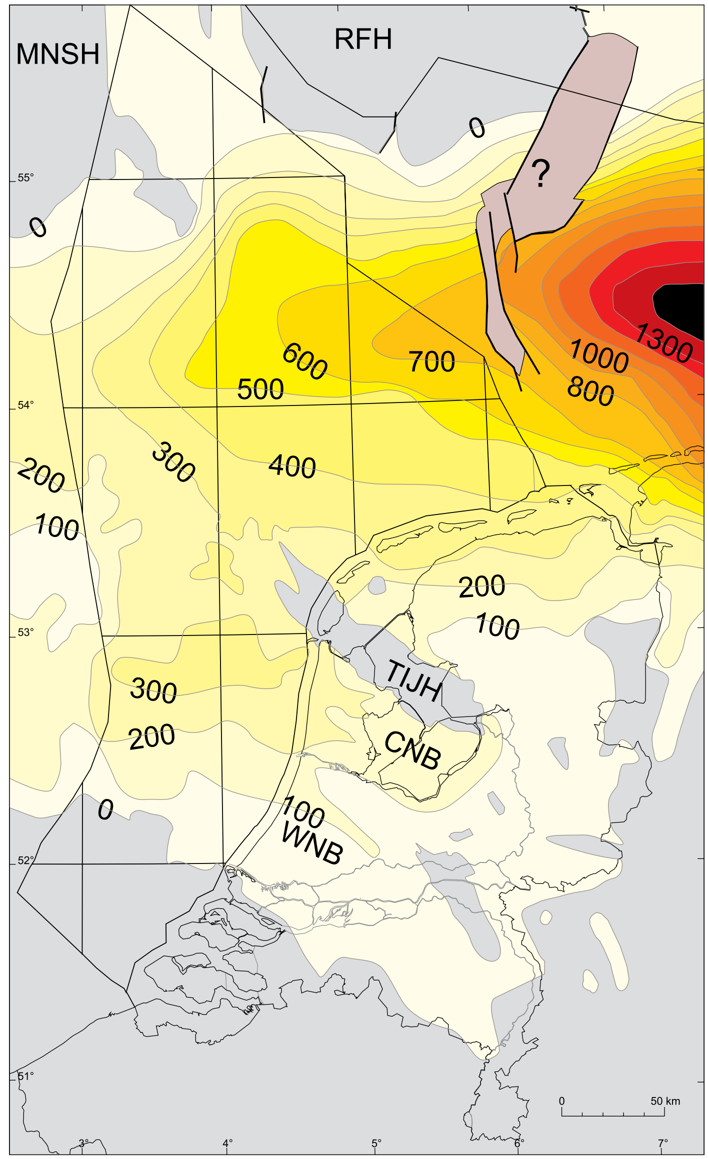


Fig. 9. Isopach map of the Upper Rotliegend Group (after Lokhorst, 1998). Contour interval is 100 m. Abbreviations as in Fig. 3. CNB: Central Netherlands Basin; WNB: West Netherlands Basin.

Lower Rotliegend Group

The Lower Rotliegend Group occurs in distinct areas, namely the Ems Low, the Dutch Central Graben and the Horn Graben (Fig. 7). It is dealt with in more detail in the chapter 'Magmatism' (Van Bergen & Sissingh, this volume). In the Ems Low, the group is represented by the *Emmen Volcanic Formation*.

The volcanoclastics in the Ems Low represent a western extension of a large area of volcanics in Germany (Plein, 1995). They consist of red-brown to green, spilitic, basaltic volcanics, mudstones and tephra layers. Several stacked lava flows occur in the succession. The thickness of the volcanics reaches a maximum of 80 m (NITG, 2000). The group is placed in the Early Permian in Germany, based on radiometric age datings (289–291 Ma) and the lithological composition of the basalts (Plein, 1995). K/Ar ages for the Lower Rotliegend Group in the eastern Netherlands (well Drouwenermond-1), however, record a much younger volcanic pulse (Middle or Late Permian, 258±6 Ma; Sissingh, 2004). This volcanic pulse correlates with the Altmark III tectonic pulse (Fig. 4), in line with the stratigraphic position of the rocks involved below the Zechstein Group or the youngest deposits of the Upper Rotliegend (Ten Boer Member; NITG, 2000).

In the Dutch Central Graben the Lower Rotliegend Group reaches a thickness of almost 150 m (Geluk, 1997,

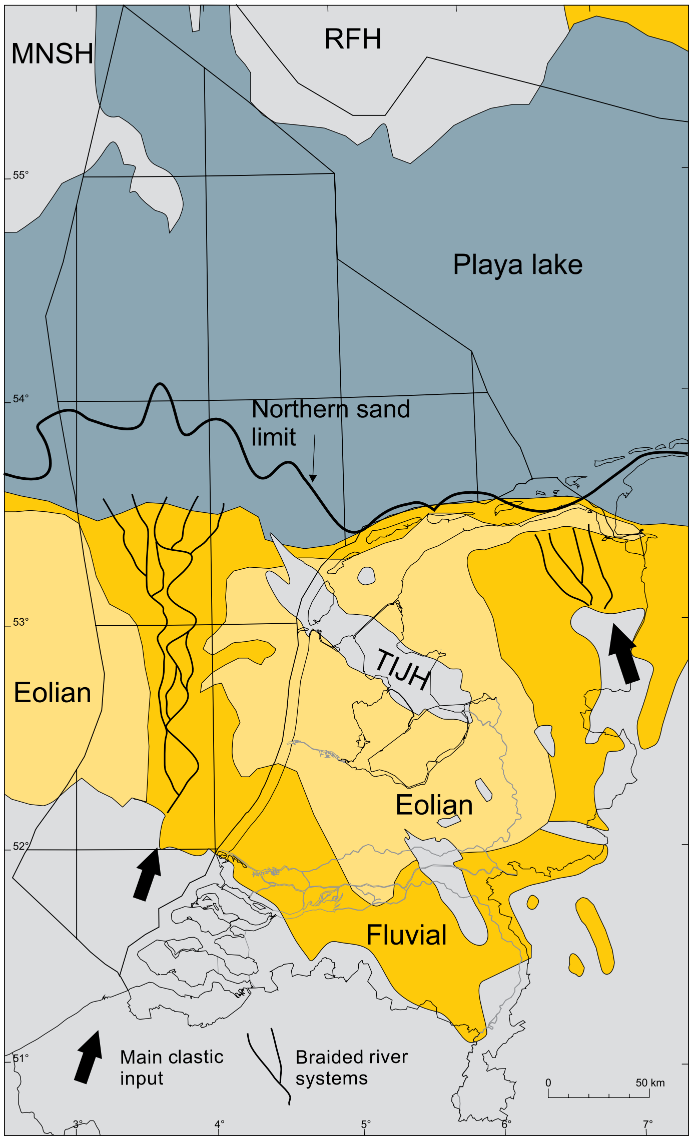


Fig. 10. Map showing the overall facies distribution at the onset of deposition of the Upper Slochteren Member (after Lokhorst, 1998). Abbreviations as in Fig. 3.

2005). The succession is lithologically similar to the Danish Karl Formation (Geluk, 2005). The distribution of the group is well-delineated towards the west, north and south by well data but remains uncertain to the east owing to a lack of such data. Based on the model of Stemmerik et al. (2000), the group could extend up to the margins of the graben. It consists of tephra and of up to several tens of metres thick, basaltic lava flows, interbedded with claystones and subordinate sandstones. Stemmerik et al. (2000) identify in the Danish North Sea three distinct volcanic pulses in the Permian, the first (300–288 Ma) and second phases (281–276 Ma), both of andesitic and rhyolitic composition and affecting the western part of the Central North Sea Graben and the Horn Graben, followed by a younger, basaltic, pulse (269–261 Ma, Middle Permian) in the Danish Central Graben. Although datings from the Dutch Central Graben are lacking, it is likely that the volcaniclastics here correlate with this younger pulse.

Upper Rotliegend Group

The Upper Rotliegend Group consists of two formations, namely the Slochteren Formation comprising mainly sandstones and conglomerates, and the Silverpit Formation composed of claystones, siltstones and evaporites. The formations are each other's lateral equivalents; their transition occurs in a relatively narrow zone in the north of the country (Fig. 8). The Upper Rotliegend reaches its greatest thickness, over 700 m, in the northern offshore, thinning rapidly both to the south and north (Fig. 9). The oldest sediments of the group occur in the northern offshore from where the sedimentation area gradually expanded southward (Verdier, 1996). The local absence of the group is due to post-depositional erosion on the Texel-IJsselmeer High and nearby smaller highs (Van Wijhe et al., 1980; RGD, 1991a, b, 1993; Rijkers & Geluk, 1996), and to non-deposition in most other areas (NITG, 1998; Geluk, 1999).

There is strong evidence for syn-depositional fault movements in the Upper Rotliegend Group. In western parts of the Central Netherlands Basin, the thickness varies across major faults (RGD, 1993). It is likely that also the northern part of the West Netherlands Basin formed a fault-bounded depression, where up to 100-m-thick eolian sands occur, although the thickness on the Zandvoort Ridge remains uncertain (see discussion in Geluk et al., 1996: p. 61–62). Other areas where faulting occurred are the eastern Netherlands (NITG, 1998), the area north of the Texel-IJsselmeer High (RGD, 1991a, b) and the Lauwerszee Trough (RGD, 1995). Thickness variations across faults are up to 25 m.

The *Silverpit Formation* comprises siltstones, claystones and evaporites. In the Dutch Central Graben, tephra layers occasionally occur in its lowermost part. The formation reaches a thickness of over 700 m in the northern offshore area and was deposited in a playa lake. It is subdivided into three members, the Lower and Upper Silverpit Claystone and the Silverpit Evaporite Member. The evaporites comprise mainly rock salt and subordinate anhydrite, interbedded with clay-siltstones. The individual rock-salt layers form good correlation horizons throughout the Southern Permian Basin. Their thickness reaches up to 50 m. In northern Germany and part of the adjacent Netherlands offshore these layers have been mobilized into salt pillows and in Germany even into diapirs during later geological history. The salt layers are not composed of pure halite, but contain abundant thin claystone beds. The formation interfingers southward with the Slochteren Formation; sev-

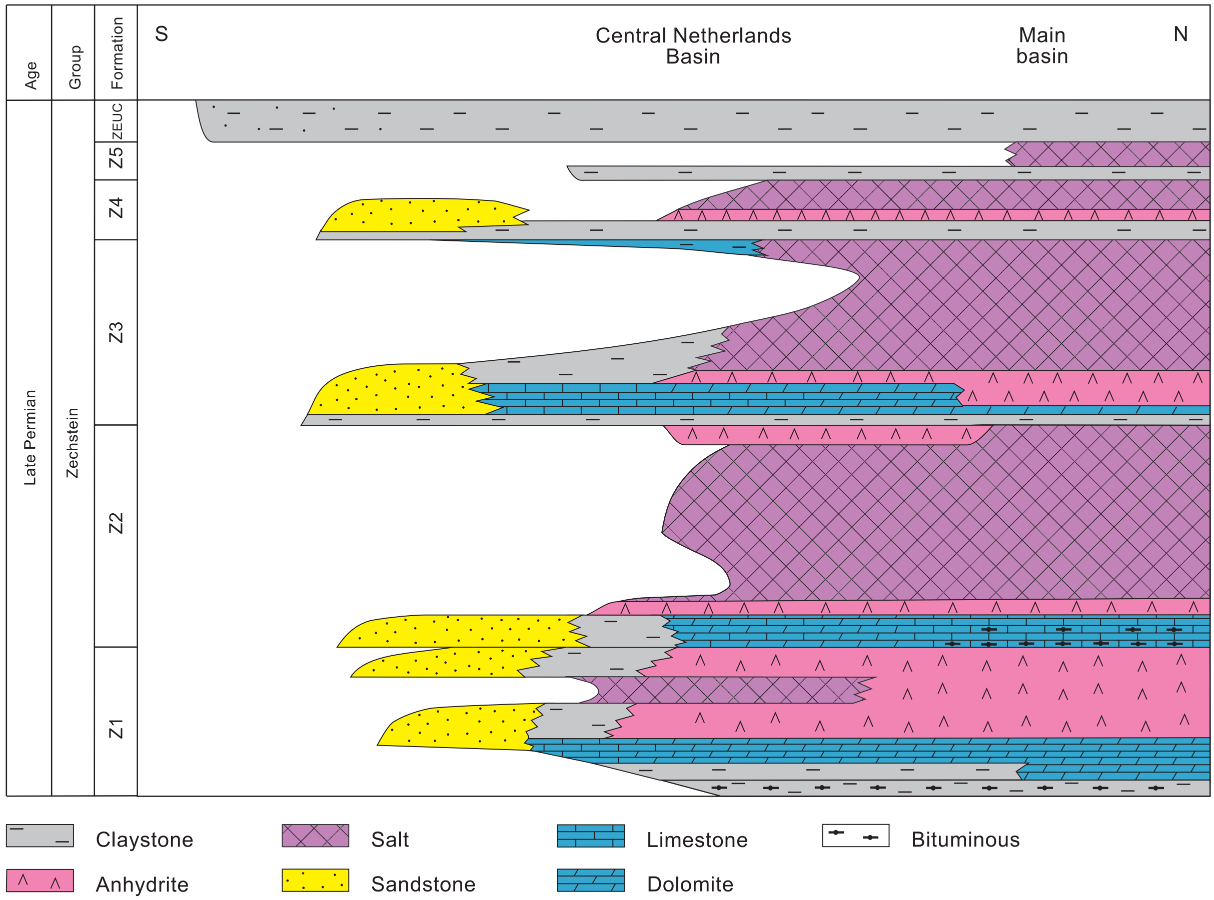


Fig. 11. Stratigraphic diagram of the Zechstein Group. In the Netherlands this group comprises five evaporite cycles (Z1-Z5) of formation rank. The Zechstein Upper Claystone Formation (ZEUC) covers these deposits unconformably. The subdivision of the formations in members reflects their dominant lithology: Z1 Carbonate, Z1 Salt, etc. (after Van Adrichem Boogaert & Kouwe, 1994).

eral tongues of the Silverpit Formation reach far to the south (Ameland and Ten Boer members).

The *Slochteren Formation* comprises conglomerates and sandstones of fluvial and eolian origin. These facies can be distinguished in cores and on dipmeter logs. Areas with predominantly eolian or fluvial deposits can be identified (Fig. 10). The fluvial systems were orientated in a S-N direction, at right angles to the basin axis. A minor clastic influx came from the Mid North Sea High. One fluvial area is situated in the western offshore, the other in the eastern onshore. Eolian deposits occur in-between these fluvial systems and to the west. Detailed studies indicate, that these facies are interbedded in a complex way, and shifted during deposition (George & Berry, 1994; Verdier, 1996); rapid vertical and lateral facies alternations occur even on a metre-scale. Eolian sandstones dominate the middle part of the formation (NITG, 2000). Within the eolian deposits a further distinction can be made between dunes, which occur predominantly in the southern areas, and damp sand flats near the playa lake (Glennie, 1998).

In the zone where the Slochteren and Silverpit formations interfinger, the former splits up into two main sandstone members, the Lower and Upper Slochteren, separated by clay- and siltstones of the Ameland Member. Eventually, both Slochteren members grade northward into siltstones and claystones of the Silverpit Formation. The stratigraphic position of the northernmost sandstone unit varies. It is in the Lower Slochteren north of the two fluvial systems and in the Upper Slochteren in the area in-between. Some isolated feather-edge sandstones and conglomerates have been encountered at the base of the Upper Rotliegend Group in the playa lake. Locally in Friesland and in the UK southern North Sea (Cameron et al., 1992) sandstones occur between the Ten Boer Member and the transgressive base of the Zechstein (Akkrum sandstone).

Zechstein Group

The Zechstein Group comprises five evaporite cycles (Z1-Z5), each of formation rank (Van Adrichem Boogaert & Kouwe, 1994). The Zechstein Upper Claystone Formation

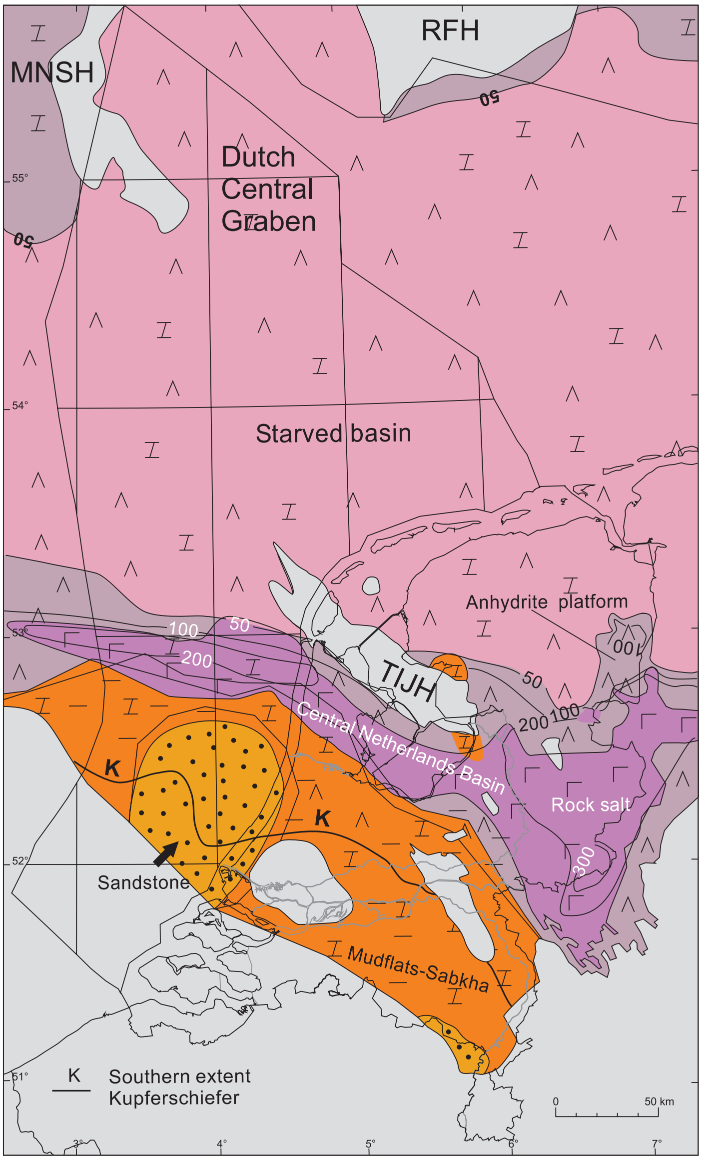


Fig. 12. Facies and isopach map (in metres) of the Z1 (Werra) Formation. The greatest thickness, over 400 m, occurs in the anhydrite platform in the east of the Netherlands in conjunction with syn-sedimentary faulting. Rock-salt deposition occurred in the area of the Central Netherlands Basin; sandstones were shed into the basin in the western offshore (arrow). In the main basin a condensed succession of carbonates and evaporites was deposited under starved conditions (after Geluk et al., 1996, 1997; Johnson et al., 1994; Baldschuhn et al., 2001). Abbreviations as in Fig. 3; lithological symbols as in Fig. 11.

covers these cycles unconformably (Fig. 11). The lithostratigraphy of the Zechstein closely approaches the concepts of genetic sequence stratigraphy (Galloway, 1989); three formation boundaries are picked on maximum flooding surfaces (base Z1: Coppershale Member; base Z3: Grey Salt Clay Member; base Z4: Red Salt Clay Member). These units form important correlation markers between the basin and basin-fringe deposits within the Southern Permian Basin (Geluk et al., 1996, 1997; Taylor, 1998). With respect to the distribution of the Upper Rotliegend, minor overstepping of the basin margins took place during the Zechstein (Geluk et al., 1996). The depositional thickness of the Zechstein Group increases from less than 50 m in the southern Netherlands, to over 1200 m in the northern offshore.

Deposition of the *Z1 (Werra) Formation* started with the Coppershale or Kupferschiefer, a 0.5-m-thick, finely laminated claystone with a total organic-carbon content of up to 5% (Lokhorst, 1998). It was deposited in most of the Netherlands, with the exception of the southern onshore area (Fig. 12). The formation further comprises the Z1 Carbonate, Z1 Anhydrite and Z1 Salt. The thickest occurrence, up to 500 m, was in the eastern part of the Netherlands where an anhydrite platform developed. The formation onlaps onto the London-Brabant Massif. In the main basin the formation has a constant thickness of some 50 m. The N-S orientation of the anhydrite platform in the east, parallel to the distribution of the Lower Rotliegend volcanics, points to a structural control on the outline of this platform (cf. Figs 7, 12).

The platform and slope deposits of the Z1 Carbonate consist of up to 200 m of marls and carbonates, in contrast to the 8 to 10 m of carbonate in the sediment-starved basinal setting (Fig. 13; Geluk, 2000). A carbonate platform was also present along the Mid North Sea and Ringkøbing-Fyn High (Taylor, 1998). Small carbonate build-ups have been encountered in the southern part of the country and adjacent parts of Germany (Visser, 1955; Teichmüller, 1957; Füchtbauer, 1980). On the anhydrite platform a series of fault-bounded depressions formed where locally up to 300 m of rock salt was deposited (Fig. 14). Potassium–magnesium salts in the Z1 occur only in the eastern Netherlands (NITG, 1998). South of the Central Netherlands Basin, the formation is composed mainly of claystones; in the offshore area medium to fine-grained fluvial sandstones occur. From the Texel-IJsselmeer High local sourcing of fine-grained siliciclastic material took place (Van Adrichem Boogaert & Burgers, 1983).

The *Z2 (Stassfurt) Formation* comprises a basal carbonate unit, the Z2 Carbonate, followed by the Z2 Basal Anhydrite and Z2 Salt. In the southern onshore areas anhydrite-bearing claystones represent this formation, while local sandstones occur in the western offshore (Fig. 15; Geluk et al., 1997). The formation is less than 50 m thick in the southern Netherlands and more than 700 m in the northern offshore. The southern limit of the Z2 Carbonate (or Main Dolomite) lies considerably more to the north compared to the Z1 Carbonate (cf. Figs 13, 16). Within the Z2 Carbonate, three facies realms can be identified: platform, slope and basin (Van de Sande et al., 1996; Geluk, 2000). The basinal deposits comprise 8 to 12-m-thick, dark-coloured, bituminous, finely laminated carbonates, known as the Stinkkalk (Fetid limestone). These rocks have source rock potential, with a total organic-carbon content of up to 1.2% (Lokhorst, 1998). In the vicinity of the carbonate platforms, slumps and turbidites of displaced shelf deposits have been identified (Amiri-Garoussi & Taylor, 1992; Van de Sande et al., 1996). The slope facies consists of light-coloured limestones and dolomites, and of platform sediments redeposited by mass flows. In the eastern Netherlands this facies reaches a thickness of over 200 m (Van de Sande et al., 1996). The platform facies comprises a complex of oolitic, pelletoidal, bioclastic and pisolitic pack- and grainstones and finely laminated wackestones (Clark, 1986; Van der Baan, 1990; Strohmenger et al., 1996). It reflects a wide range of depositional settings: sabkha, tidal flats, algal shoals, lagoons and ooid bars and shoals (Strohmenger et al., 1996). 3D-seismic interpretation revealed a complex outline of the platform, including isolated off-platform highs (Van de Sande et al., 1996). The shape of the Z1 Anhydrite platform controlled to a great extent the facies distribution of the Z2 Carbonate (cf. Figs 12, 14). The carbonates grade southward into anhydritic clay-siltstones, and in the eastern Netherlands, rock salt (NITG, 2000). In the eastern Netherlands, the Z2 Carbonate platform complex measures some 50 to 70 km wide in a N–S direction, whereas in the western offshore area this is much less, between 10 and 30 km.

The topography of the Z2 Carbonate controlled the thickness and distribution of the overlying units; the thickest rock salts were deposited in the carbonate basin, and thin salt in platform areas. The Z2 Salt is over 600 m thick in the basin and is mainly composed of halite (> 95%). Three stages of deposition can be identified within this salt (Geluk, 1995; Geluk et al., 2000). During the first two stages the initial topography still persisted, and platform halites graded northward into deeper-water salt complexes (intercalated halite, polyhalite and carnallite), whereas the third stage of salt deposition filled most of the remaining relief of the basin. At the top, a regionally well-developed potassium–magnesium salt unit occurs (Fig. 17). The Z2 Salt forms the main regional top seal for the Upper Rotliegend gas reservoirs, and is also the main diapiric salt. Because of the extensive salt movement, the units overlying the Z2 Salt have become strongly deformed, which complicates the reconstruction of their primary thickness and composition.

The *Z3 (Leine) Formation* comprises the Grey Salt Clay, Z3 Carbonate, Z3 Main Anhydrite and Z3 Salt. The basal member of the formation, the Grey Salt Clay, forms an important regional marker within the Zechstein. Its thickness varies from 5 to 10 m. The facies pattern of the Z3 Carbonate is not as well-developed as in the Z2 Carbon-

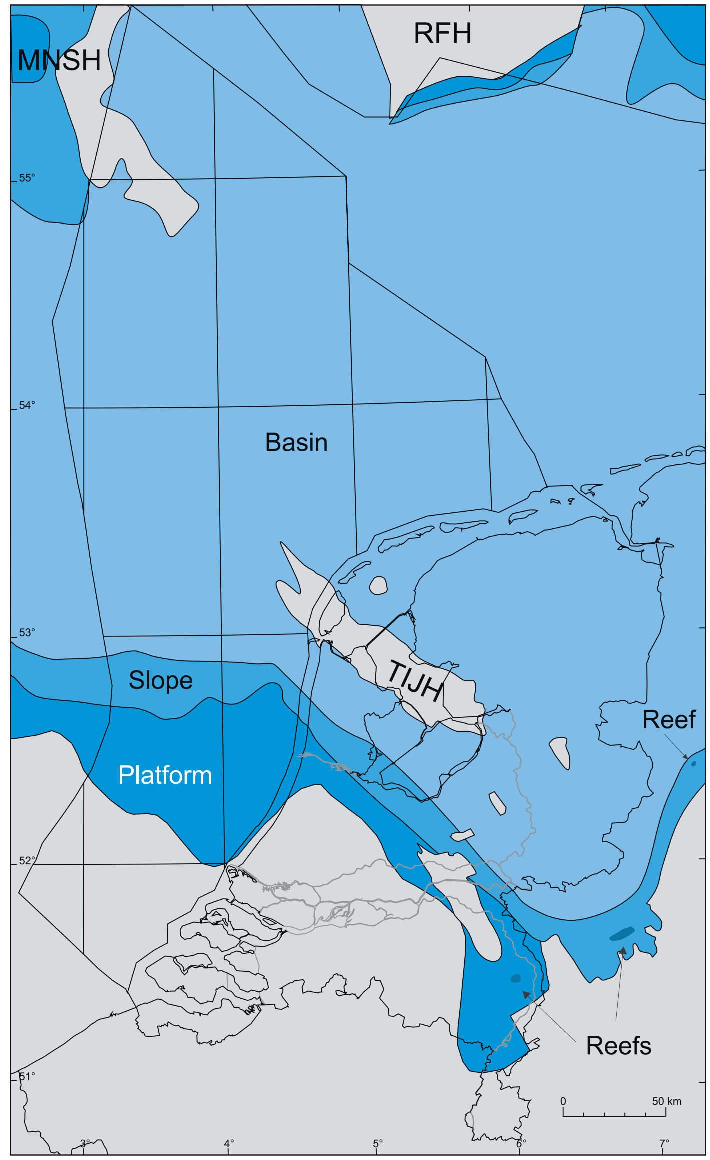


Fig. 13. Facies map of the Z1 Carbonate Member. Carbonate platforms fringed the basin both to the north and to the south. Reefs have been identified at the southern margin in Germany and the Netherlands (after Johnson et al., 1994; Geluk, 2000). Abbreviations as in Fig. 3.

ate (Geluk, 2000), reflecting the levelling out of the paleotopography by deposition of the Z2 Salt (Figs 18, 19). In the basin, the carbonate comprises a dark-coloured limestone, which is only a few metres thick. The slope facies comprises laminated and bioturbated carbonate mudstones and silty dolomites, with a thickness of up to 40 m. The platform facies consists dominantly of grey microcristalline dolomites and algal boundstones. These are considered by Taylor (1998) to represent predominantly shallow, quiet-water, possibly lagoonal deposits. Oolitic and bioclastic grainstones locally occur in the area ad-

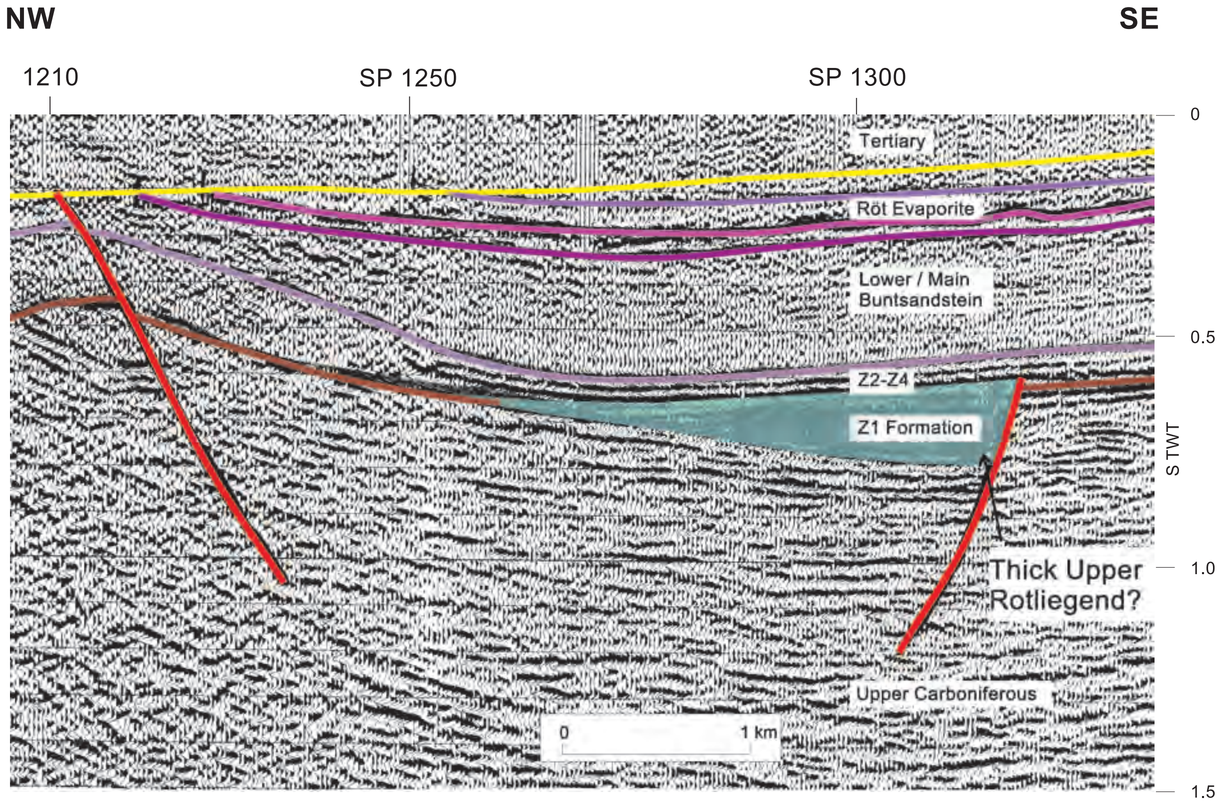


Fig. 14. Example of an anhydrite and halite-filled Zechstein rift structure in the easternmost part of the Central Netherlands Basin in east Gelderland. The Z1 (Werra) Formation shows a distinct increase in thickness, whereas younger Zechstein formations are not influenced by the faulting (after Geluk, 1999). The halfgraben is assumed to have formed part of the paleotopography of the Base Permian Unconformity, thus containing thicker Upper Rotliegend deposits than on the adjacent flanks (Geluk & Mijnlieff, 2001).

jacent to the slope (Van Adrichem Boogaert & Burgers, 1983; Baird, 1993) and on the landward margin of the platform. Off-platform shoals have been encountered on the Groningen High (Fig. 19). In the western offshore area, the carbonate grades into fluvial sandstones (Fig. 18; Geluk et al., 1997).

The Z3 Main Anhydrite was deposited only on the northern part of the Z3 carbonate platform and slope, and in the basin. The thickness of this anhydrite amounts to up to 100 m, showing complex and rapid changes (NITG, 2000). Movement of the underlying Z2 Salt caused the breaking-up of this member into large slabs, known as rafts or floaters, which became embedded in the salt and present a drilling hazard, potentially causing losses of drilling fluids or influx of high-pressured brines or gas. The Z3 Salt is composed of a basal part consisting of halite, and an upper part comprising two thick potassium–magnesium salt layers. The salts encountered include beds up to 10 m thick of the rare bischofite ($MgCl_2 \cdot 6H_2O$), one of the most soluble salts on earth, besides kieserite, carnallite and sylvite (Coelewij et al., 1978). These salts are present only in the north-eastern onshore and north-western offshore. The thickness of the Z3 Salt reaches 300 to 400 m.

The *Z4 (Aller) Formation* is composed of a basal claystone, the Red Salt Clay, followed by the thin Z4 Pegmatite Anhydrite and the Z4 Salt. The lower two members have a wide distribution, whereas the Z4 Salt is found only in the depocentres or as isolated occurrences in halfgrabens (Geluk, 1999). The thickness of the salt reaches 150 m. Potassium-magnesium salts occur in the middle part of the salt; its upper part is composed of an alternation of halite and claystone. Along the southern basin margin, the formation comprises anhydritic claystones, deposited in sabkhas, and medium to coarse-grained fluvial sandstones in the western offshore area (Fig. 20). These sandstones are known in the UK as the Hewett Sandstone (Geluk et al., 1996).

The occurrence of the *Z5 (Ohre) Formation* is limited to the north-east of the country and to the north-western offshore, outlining the depocentres towards the end of Zechstein sedimentation (Fig. 5). It comprises a basal claystone with a thickness of several metres followed by up to 15 m of halite. Younger Zechstein salts (Z6 and Z7) have been reported from north-west Germany (Best, 1989). They are absent in the Netherlands, possibly as a result of non-deposition.

The *Zechstein Upper Claystone Formation* occurs throughout the country and accumulated disconformably upon older rocks of the Zechstein and even the Upper Rotliegend Group (Fig. 5). Based upon its palynological content, the formation has been included in the Zechstein (Van Adrichem Boogaert & Kouwe, 1994). The formation is composed of red and grey anhydritic claystones and sandstones, deposited in a lacustrine to mudflat setting. Its thickness varies from 10 to 50 m.

Paleogeography

Deposition during the Permian took place in the large Southern Permian Basin, at a paleolatitude around 10° N, and north of the Variscan Mountains. These mountains prevented humid air masses reaching the area from the Tethys Ocean south of these mountains (Glennie, 1998), and consequently the climate was arid. In Early Permian times, widespread intrusive activity occurred in the Netherlands (Eigenfeld & Eigenfeld-Mende, 1986; Ziegler, 1990; Van Bergen & Sissingh, this volume). During the Middle Permian, local extrusion volcanism (Lower Rotliegend Group) occurred at different times in relation to wrenching along deep-seated faults in the Ems Low and the Dutch Central Graben (Fig. 7). In the Dutch Central Graben, minor amounts of clastic sediments were deposited. This extrusive volcanism is younger than in Germany and Poland, where it already started in Late Carboniferous to Early Permian times (Plein, 1995). The pulses are thought to accompany the opening of the Horn and the Central North Sea Graben (Glennie, 1997; Stemmerik et al., 2000). Prior to the deposition of the Upper Rotliegend Group, erosion resulted in highly differentiated and locally deep truncation of the Carboniferous. The paleorelief was not completely levelled out (Van der Baan, 1990; Van de Sande et al., 1996); differential weathering and faulting were responsible for local relief of up to 25 m (Fig. 6; Geluk & Mijnlieff, 2001). In the north-east of the country, the area of Middle Permian volcanic activity still had a positive relief, and became covered only by the Zechstein transgression or, locally, the youngest Upper Rotliegend sediments (NITG, 2000).

Sedimentation of the Upper Rotliegend started in Poland and north-east Germany during the Early Permian (Upper Rotliegend I; Plein, 1995). The sedimentation area expanded with time and reached the Netherlands in the Middle Permian in the northern offshore. Deposition of the Upper Rotliegend did not keep pace with basin subsidence, resulting in a large inland depression below sea level (Glennie & Buller, 1983). In the central part of this area, a large playa lake developed where claystones and rock salt were deposited (Fig. 1).

Important fluvial feeder systems were located in the western offshore and in the Ems Low (Fig. 10). The Variscan Mountains formed the main source areas of clastics. The systems were formed by wadis with an episodic run-off. With prevailing north-east trade winds, the sand was deflated from the fluvial deposits during dry periods and accumulated down-wind as dunes or, close to the playa lake, as damp sand flats (Glennie, 1998). The climate during deposition of the Upper Rotliegend was characterized by an alternation of wet and dry periods, caused by Milankovitch cycles (Yang & Baumfalk, 1994; George &

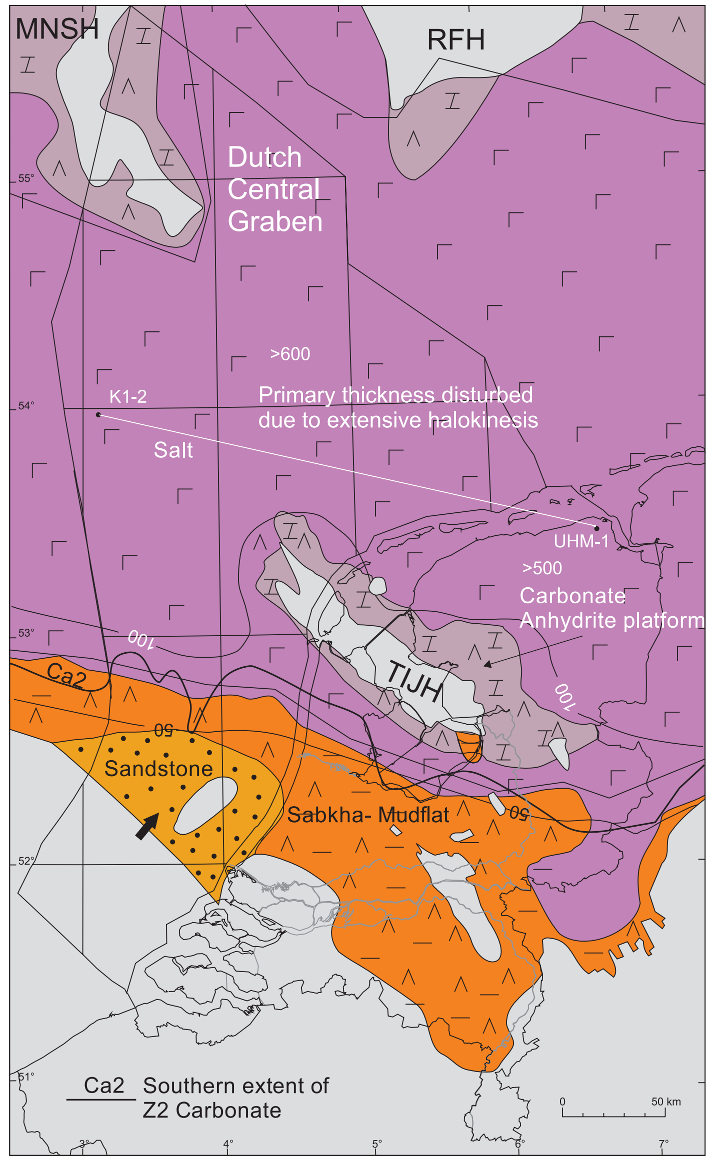


Fig. 15. Facies and isopach map (in metres) of the Z2 (Stassfurt) Formation. A thick succession of rock salt, over 600 m, was deposited in the centre of the Southern Permian Basin. In the southern onshore, extensive sabkha-mudflats developed. Sandstones were shed into the basin in the western offshore (after Johnson et al., 1994; Geluk et al., 1996, 1997; Baldschuhn et al., 2001). Abbreviations as in Fig. 3. UHM-1: Uithuizermeeden-1 (Fig. 17).

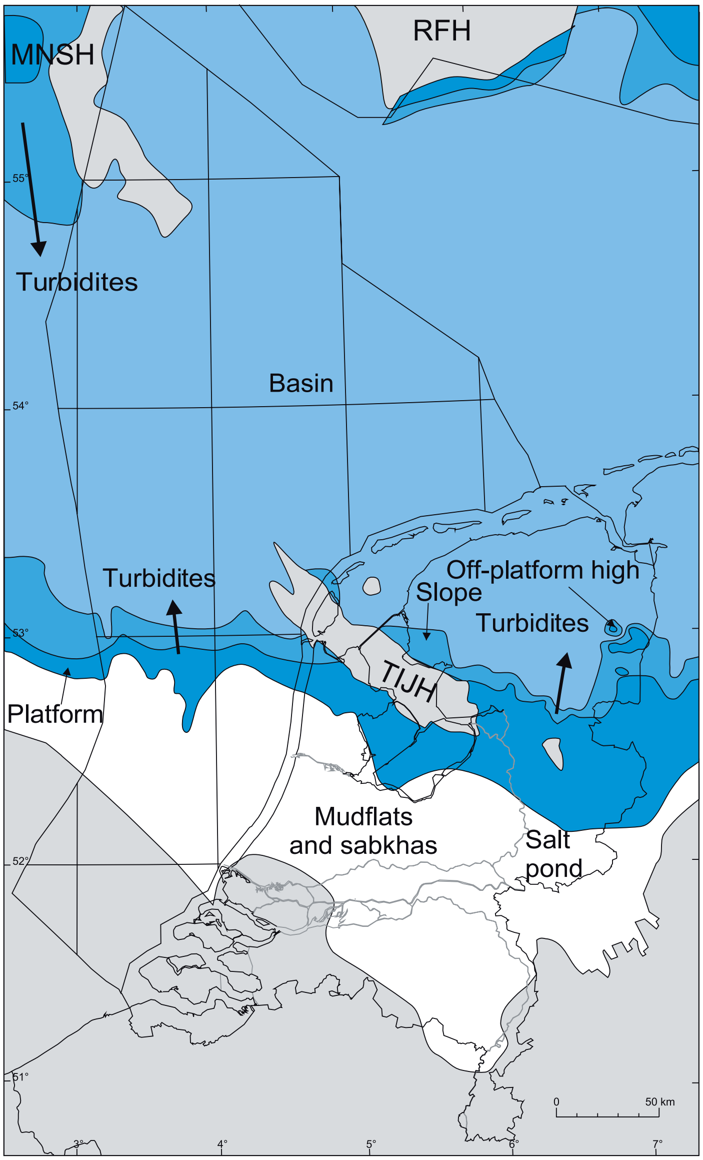


Fig. 16. Facies map of the Z2 Carbonate Member. Well-developed platform facies, including off-platform highs, were situated in the eastern Netherlands and to the north of the basin. The basinal deposits comprise organic-rich carbonates. Turbidites were shed from the platforms into the basin (after Johnson et al., 1994; Geluk, 2000). Abbreviations as in Fig. 3.

Berry, 1994). This cyclicity caused a periodic expansion and retreat of the lacustrine facies belts, and a complex alternation of fluvial and eolian deposition at the margins of the lake.

In Late Permian times, rifting created a connection between the Southern Permian Basin and the Barents Sea (Ziegler, 1990). A widespread reworking and redeposition of dune sands took place at the end of the Upper Rotliegend deposition by increased fluvial activity (K.W. Glennie, personal communication). Under the prevailing dry climate, widespread cyclic deposition of evaporites took place. The controlling mechanism behind the cyclicity is thought to have been of glacio-eustatic origin (Ziegler, 1990). Five Zechstein cycles have been identified in the Netherlands. Transgressions mark the bases of these cycles. In the lower cycles (Z1 to Z3) normal-marine conditions were established throughout the Southern Permian Basin. Clastic influx was pushed back to the basin margins (Geluk et al., 1996). During higher cycles (Z4 and Z5) no carbonates were deposited; the conditions were permanently hypersaline. On the whole, the carbonates and evaporites were laid down as a series of prograding and aggrading sigmoidally shaped bodies, where the facies belts and the area of maximum thickness shifted with time towards the north (cf. Figs 13, 16, 19 for the carbonate units).

Following the initial transgression, a starved basin, up to 200 m deep, was established, where anoxic sediments of the Kupferschiefer were laid down (Sweeney et al., 1987). Normal-marine, oxic conditions were established in the basin afterwards. The interaction of an inclined basin topography, minor faulting and sea-level rise led to the establishment of a carbonate platform over the southern onshore and western offshore areas, where locally small carbonate build-ups formed (Figs 12, 13). Around the London-Brabant Massif extensive mudflats and sabkhas were present, whereas in the western offshore sands were deposited under fluvial to estuarine conditions (Chris Elliot, personal communication). The geometry of the carbonate deposits in the western offshore suggests a relief of up to 200 m. After this transgression a lowering of the sea level followed, during which the connection with the Barents Sea became restricted and evaporites were deposited. At the basinward side of this carbonate platform a large anhydrite platform developed. In the shallow, warm water, anhydrite accumulated rapidly, in contrast to slow precipitation in the basin area north of the platform (Van der Baan, 1990). The similarity between the outline of the anhydrite platform and areas of previous volcanic activity points to structural control (cf. Figs 7, 12). A combination of minor faulting and loading of the thick anhydrite deposits upon a differentiated substrate created a series of depressions in this platform area, where rock salt was deposited (Geluk et al., 1997; Geluk, 1999).

The second Zechstein transgression did not reach as far southwards as the Z1 transgression (Fig. 16). An extensive carbonate platform developed, with oolite shoals and lagoons (Figs 2, 16, 21; Van der Baan, 1990; Strohmenger et al., 1996; Van de Sande et al., 1996; Geluk, 2000). To the south, this platform graded into a complex of fluvial sandstones, mudflats, sabkhas and salt ponds (NITG, 1998). Lowering of the sea level resulted in widespread evaporite sedimentation. The basin was filled with halite, with

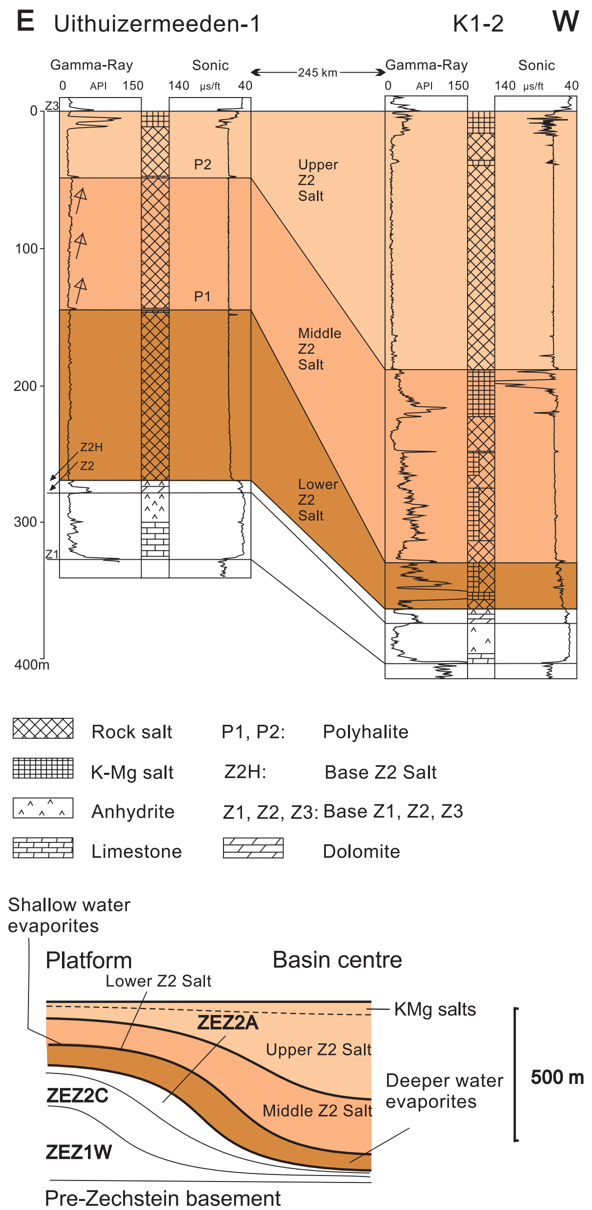


Fig. 17. Correlation of the Z2 Salt Member, and below a 2D model for the Z2 Salt sedimentation. The well Uithuizermeeden-1 reflects a platform setting, whereas the K1-2 well was situated in the central part of the basin (location on Fig. 15). The lower Z2 Salt grades from shallow-water halites into a deeper-water salt complex. Arrows in the middle Z2 Salt indicate cycles of increasing potassium content. The upper Z2 Salt filled the remaining relief in the basin. At the top of this unit, potassium-magnesium salts occur (after Geluk et al., 1997). ZEZ1W: Z1 Anhydrite; ZEZ2C: Z2 Carbonate; ZEZ2A: Basal Anhydrite.

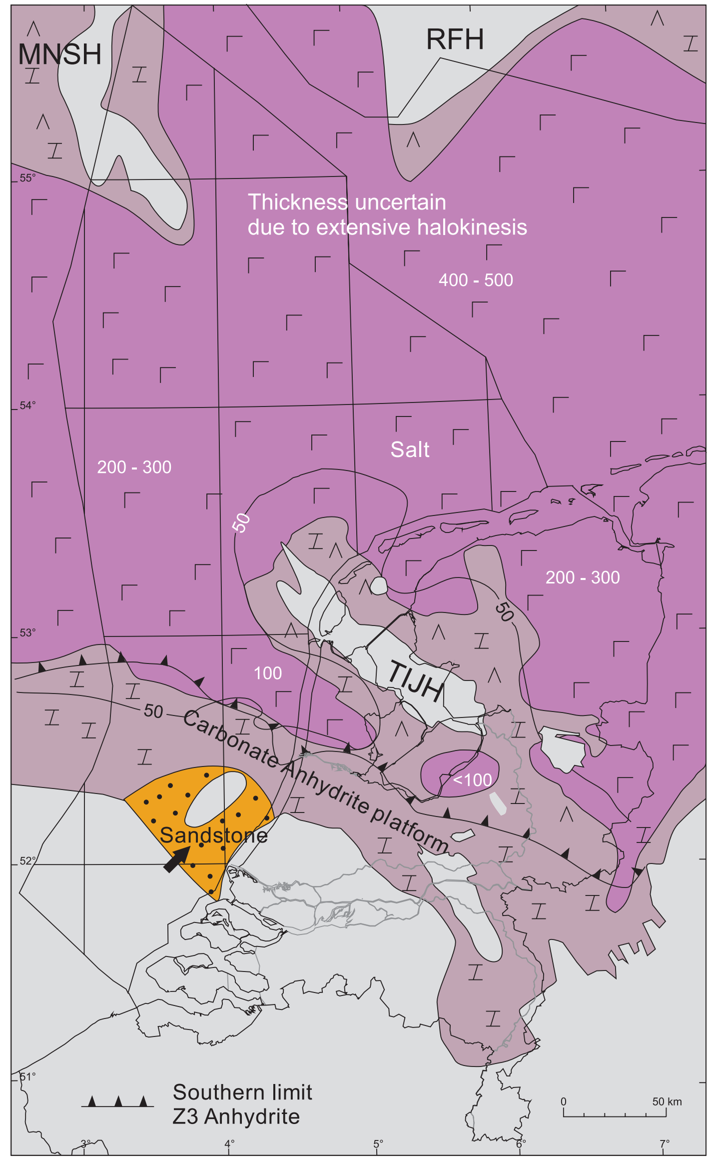


Fig. 18. Facies and isopach map (in metres) of the Z3 (Leine) Formation. This formation records the most widespread transgression, with carbonate development extending towards the margins of the basin. Simultaneously, clastic deposition continued in the western offshore. The depositional salt thickness in the central part of the Southern Permian Basin is uncertain (after Johnson et al., 1994; Geluk et al., 1997; Baldschuhn et al., 2001). Abbreviations as in Fig. 3.

a thickness up to 600 m. The salt deposition was multicyclic, triggered by minor transgressions causing salinity fluctuations (Geluk, 1995). The lower and middle salts display characteristic prograding foresets, the upper salt filled the remaining relief (Fig. 17). In contrast to Tucker (1991), who considers the entire Z2 Salt as a lowstand deposit, only the Upper Z2 Salt is considered here as such.

The third transgression records the last normal-marine salinities in the basin. This transgression reached much further southward than the previous ones and pushed fluvial deposits back to the basin margin (Figs 18, 19). The facies variations within the carbonate deposits decreased, as the relief in the basin was mostly flattened out. The re-

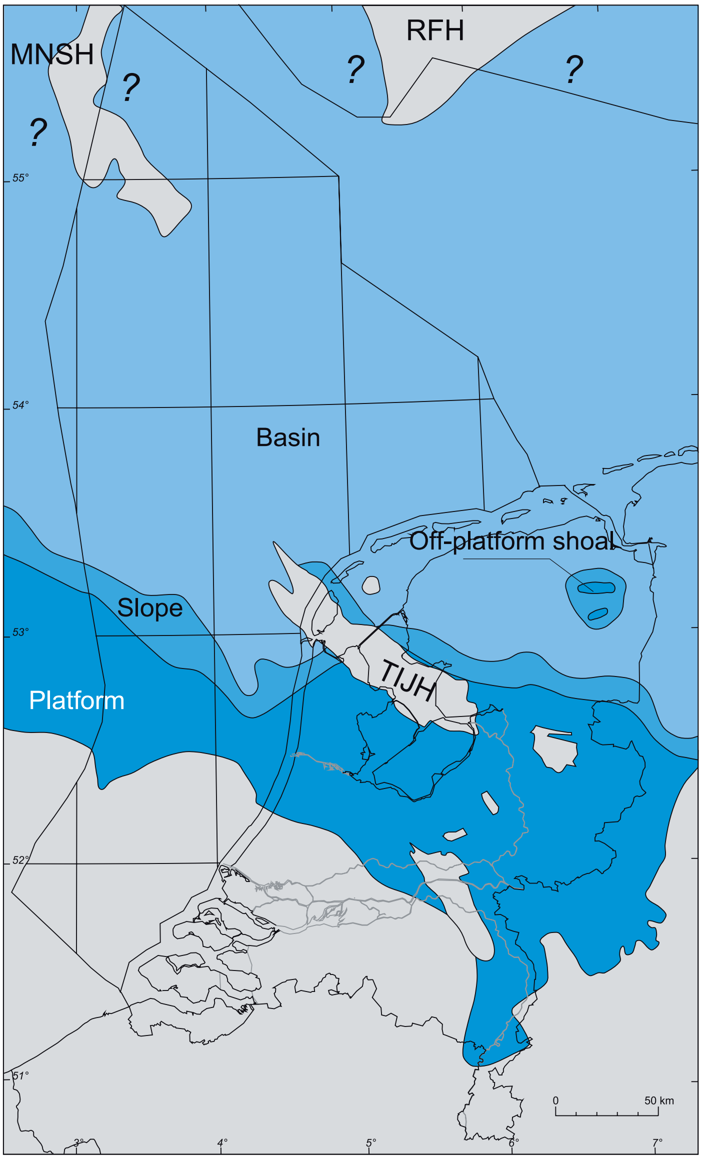


Fig. 19. Facies map of the Z3 Carbonate Member. This member has the greatest southward extension of all Zechstein carbonates. The paleogeography represents a more simple, ramp-type development than the Z2 Carbonate (Fig. 16). A shoal was situated above the Groningen High (after Geluk, 2000). Abbreviations as in Fig. 3.

lief did not exceed several tens of metres. A carbonate-anhydrite platform occupied most of the area on and around the Texel-IJsselmeer High, while carbonates dominated the southern Netherlands. The up to 10-m-thick beds of bischofite and potassium-magnesium salts in the Z3 Salt indicate that complete desiccation occurred in the north-east of the country.

From the fourth transgression onwards, conditions in the basin became permanently hypersaline (Ziegler, 1990). Claystones were deposited in hypersaline shallow-water environments (Fig. 20). The thick, massive halites and potassium–magnesium salts in the lower part of the Z4 (Aller) Formation were probably still deposited in a marine environment; in the upper part of the formation the alternation of saliferous claystone and thin halites indicates a gradual change to playa-type conditions. The brines, however, were still of marine origin. Deposition was governed by higher-order Milankovitch cycles, with halites representing the dry, and claystones the wet periods (Wagner, 1994; Geluk et al., 1997). The Z5 (Ohre) Formation was likewise deposited under playa conditions.

The higher Zechstein cycles Z6 and Z7, described by Best (1989) from north-west Germany, were not deposited in the Netherlands. Contemporaneously with the deposition of these higher cycles, uplift and erosion affected the Zechstein in the Netherlands, in some areas even the entire Zechstein (Geluk, 1999). The Zechstein Upper Claystone Formation represents a widespread development of sand- and mudflats throughout the Southern Permian Basin (Geluk et al., 1997).

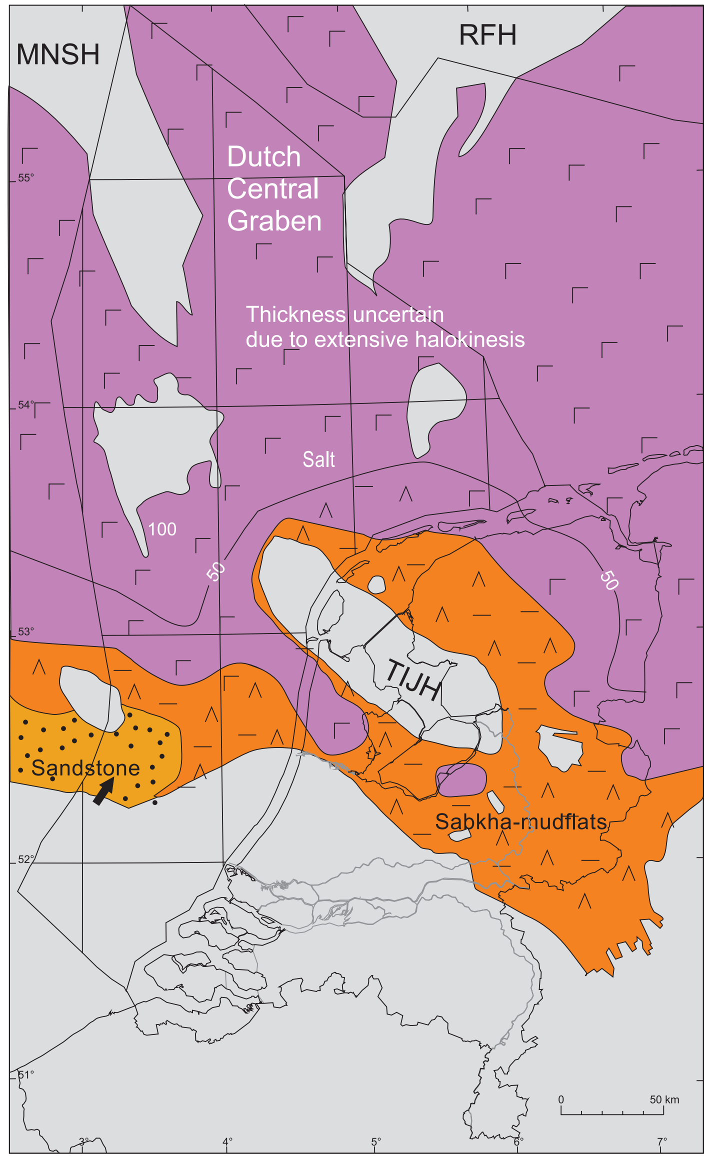
Salt movement

Zechstein rock salt greatly influenced the post-Permian structuration of the Netherlands. Because of its great thickness, in many areas up to 1000 m, and its visco-plastic behaviour during deformation, the salt acts as a major detachment zone, separating the sub-salt fault system from the faults in the overburden. The visco-plastic behaviour of the salt in combination with its low density led to the development of numerous salt structures (Fig. 22). Trusheim (1963) recognized the importance of the buoyancy forces to mobilize the salt, but recent studies point to a tectonic trigger to initiate salt movement (Remmelts, 1995, 1996). This is indicated by comparison of the orientation of the salt structures and the sub-salt structural grain. Furthermore, tectonic stress in general results in an increase of the rate of salt movement (Remmelts, 1996; Baldschuhn et al., 1998). The salt movement started on a small scale during the Early Triassic and culminated during the Jurassic in response to extensional tectonics. Compressive stresses during the Late Cretaceous laterally squeezed and strongly deformed many salt structures (Baldschuhn et al., 1998; NITG, 2000). Cenozoic tectonics resulted in renewed salt movement (Remmelts, 1996).

To understand the structural development of the Netherlands, it is important to differentiate between areas with thick salts and those without salt. The structural styles of these areas are completely different, as were their responses to the inversion tectonics (De Jager, this volume).

Apart from the overall structuration, salt movement can play a significant role on a local scale: areas of salt withdrawal provide extra accommodation space for sediments, and Remmelts (1996) showed that the process of salt flow created various types of structural traps for hydrocarbons

Fig. 20. Facies and isopach map (in metres) of the Z4 (Aller) Formation. In much of the Netherlands the formation is represented by thin sabkha-mudflat deposits. The sandstone influx into the basin shifted to the west (cf. Fig. 18). In UK stratigraphy this sandstone is known as the Hewett Sandstone, and considered as Lower Triassic (Johnson et al., 1994). The depositional salt thickness in the central part of the Southern Permian Basin is uncertain. Abbreviations as in Fig. 3.



in their vicinity within the Netherlands (turtle-back anticlines, four-way dip closures, fault–dip closures).

Economic geology

Permian deposits are of great importance to the Netherlands from an economic point of view, since over 95% of the country's natural gas reserves are trapped in Permian reservoirs (De Jager & Geluk, this volume). Besides gas accumulations, a few Permian reservoirs contain oil. In addition, Permian evaporites are important to the salt industry; apart from rock salt they contain potassium and magnesium salts (Geluk et al., this volume).

Natural gas

Permian rocks form the prime hydrocarbon exploration target in the Netherlands. Most gas accumulations are located in the clastic reservoirs of the Upper Rotliegend. Carbonates of the Zechstein form a second reservoir unit, mainly in the former platform areas. The gas in the Upper Rotliegend reservoirs stems mainly from the Upper Carboniferous coal measures (Van Wijhe et al., 1980; Glennie, 1998), with contributions in some areas (northern Netherlands, north-west offshore) from the basal Namurian hot shales (Gerling et al., 1999). Zechstein salt seals the Upper Rotliegend gas accumulations. In the marginal areas of the basin, where no Upper Rotliegend is present, or where the salt seal is thin, gas accumulated in Zechstein carbonate reservoirs, sealed by Zechstein salts or anhydrites. In these reservoirs, varying amounts of condensate and also H_2S occur (Lokhorst, 1998).

Oil

Minor amounts of oil have been encountered in the north-east of the country in the Upper Rotliegend at Midlaren and in the Zechstein in two fields (Stadskanaal and Gieterveen) as well as in one Zechstein discovery in the north-

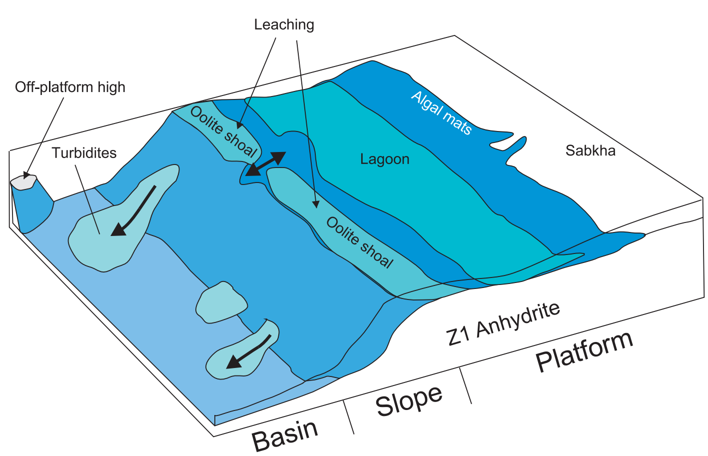


Fig. 21. Depositional model of the Z2 Carbonate. In the basin, fine-grained, organic-rich lime muds were deposited, with, in the vicinity of the carbonate slope, distal turbidites. On the slope, these fine-grained sediments intermix with coarser redeposited platform deposits. The platform itself covers a wide range of depositional environments, ranging from oolite shoals on the basinward side, via lagoonal deposits to algal mats and sabkhas on the landward side.

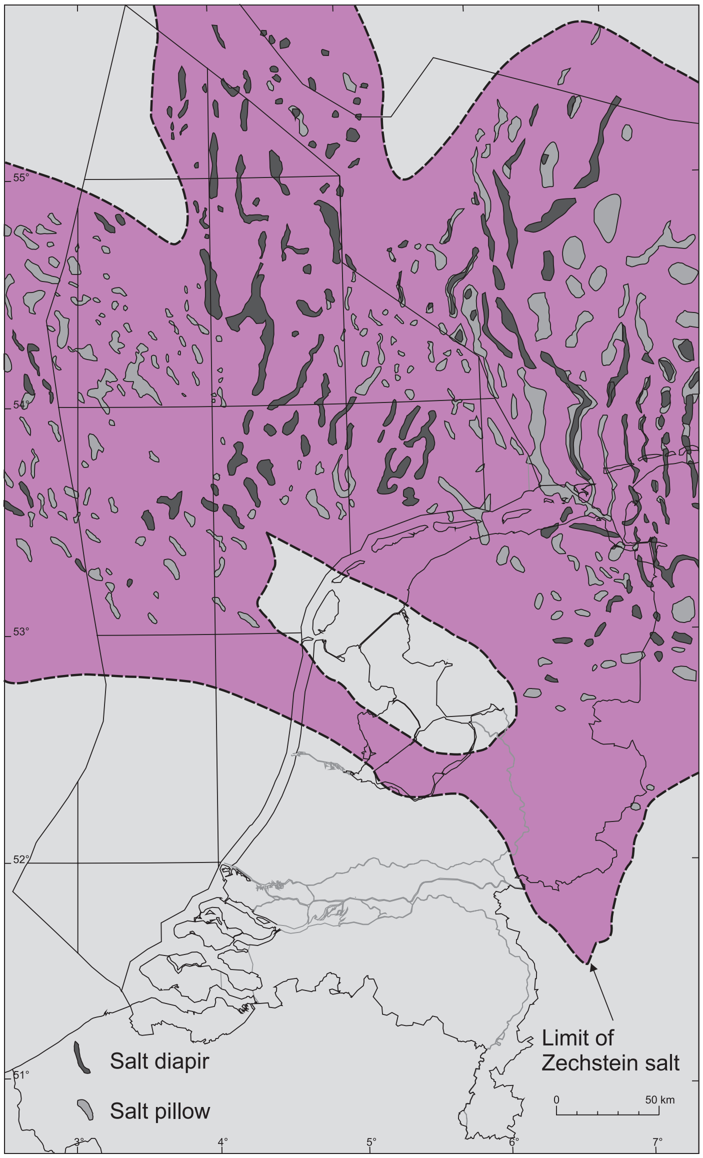


Fig. 22. Map of the Zechstein salt structures in the Netherlands (after Lokhorst, 1998). Elongated salt domes and salt walls are related to major sub-salt faults, whereas circular domes developed at intersection points of different fault trends (Remmelts, 1996).

western offshore (E13-1). The most likely source rock for this oil is the Stinkkalk, the basinal equivalent of the Z2 Carbonate (De Jager & Geluk, this volume). In the Zechstein accumulations, the Z2 Carbonate forms the reservoir, the Z2 Salt the seal.

Rock salt

Rock salt (halite) is produced by solution mining from the Z2 Salt in the northern Netherlands. In the Adolf van Nassau concession, in the province of Groningen, the salt is produced from two diapirs, which are located at relatively shallow depths. The production started in the 1960s. Salt production also takes place from deeply buried bedded salt in the province of Friesland, in the Barradeel concession. Here, the salt is produced from depths between 2500 and 3000 m, making it the deepest salt produced in the world.

Potassium–magnesium salts

Potassium–magnesium salts are produced by solution mining in the Groningen province, in the Veendam concession. Here, thick beds of various salts occur in the Z3 Salt. They include carnallite, sylvite and the rare bischofite (Coelewij et al., 1978).

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Triassic

M.C. Geluk

ABSTRACT

The Lower and Upper Germanic Trias groups in the Netherlands are separated by the Base Solling or Hardegsen Unconformity. The lower group (latest Permian–Olenekian, up to 800 m thick) is made up mainly of fine-grained siliciclastic deposits with sandstone and oolite intercalations. The fines represent lacustrine sediments; the sandstones are fluvial and eolian. The source area of these clastics was the Variscan Mountains to the south. The distribution and thickness of the sandy sediments were controlled by rift tectonics (Hardegsen phase). The Upper Germanic Trias Group (Olenekian–Norian, up to 1750 m thick) comprises lacustrine, brackish-water and marine, fine-grained siliciclastics, carbonates and evaporites. Sandstones are rare. The Early Kimmerian rift tectonics strongly influenced the thickness and distribution of these sediments. The Permian basin configuration was initially maintained, but was modified progressively by the rifting. In Middle Triassic times a connection was established between the Permo-Triassic basin and the Tethys Ocean, resulting in the deposition of shallow-marine carbonates and evaporites. Differential subsidence started in the West Netherlands Basin and Roer Valley Graben from the Anisian onwards, but in the Dutch Central Graben it had already begun during the Permian. Uplift of the Fennoscandian Shield in the Early Ladinian resulted in clastic supply from the north-east. Triassic rocks form the second-largest hydrocarbon reservoir in the Netherlands. Rock salt from the Triassic is produced in the east of the country by solution mining. A single quarry in the east produces limestone and dolomite from the Lower Muschelkalk.

Keywords: Netherlands, Buntsandstein, Muschelkalk, Keuper, Kimmerian rifting, carbonates, evaporites, sandstones

Introduction

The Triassic rocks in the Netherlands represent part of the post-Variscan sedimentary mega-cycle, which lasted until the Middle Jurassic. They host natural gas and oil, and form the country's second-largest hydrocarbon reservoir (De Jager & Geluk, this volume). Triassic rock salt (Geluk et al., this volume) and carbonates are exploited in the eastern Netherlands.

During the Early Triassic, sedimentation continued in an area essentially similar to the Southern Permian Basin, but under continental conditions (Fig. 1). In the course of the Triassic, however, Early Kimmerian extensional tectonics broke the basin up in a number of fault-bounded depocentres, and triggered wide-spread mobilization of Zechstein salt (De Jager, this volume). Connections with the Tethys through the Silesian-Moravian Gateway in southern Poland and the more westerly Burgundy gate resulted in a wide-spread deposition of marine carbonates of the Muschelkalk during the Middle Triassic (Fig. 2). Uplift of Fennoscandia during the Late Triassic resulted in fluvial systems building out towards the south (Fig. 3).

Lower Triassic rocks occur in the subsurface of most of the country; Middle and Upper Triassic rocks have a more restricted occurrence and are confined mainly to the Jurassic–Cretaceous rift basins. The depth of the base of the Triassic varies from positions close to the surface in the eastern and southern Netherlands to over 7 km in the Dutch Central Graben (Fig. 4).

The Triassic rests conformably on the Permian in most of the Netherlands. Locally, at the southern margin of the basin, it oversteps onto the Carboniferous. The Triassic rocks are of epicontinental character and were deposited in eolian, fluvial, lacustrine, paralic and shallow-marine environments. The thickness of the Triassic varies strongly as a result of a highly differentiated intra-Triassic subsidence pattern and significant post-Triassic uplift and erosion; in the Dutch Central Graben up to 2500 m of Triassic rocks are present, in the Ems Low 2000 m and in the other basins around 1000 m. On many highs bordering these basins only a reduced succession of Triassic rocks remains of several hundred metres in thickness. Until Middle Triassic times the source areas for the clastics were the Variscan Mountains to the south. Triassic rocks record the onset of the Mesozoic rifting, which accompanied the breaking-up of Pangea. During Late Triassic times the Fennoscandian Shield formed the main sediment source.

Stratigraphy

The Triassic rocks are subdivided into two groups (Table 1; Van Adrichem Boogaert & Kouwe, 1994):

– the Lower Germanic Trias Group (latest Permian–Olenekian), comprising mainly fine-grained clastic deposits with sandstone and oolite intercalations, and consisting at the southern basin margin predominantly of sandstones;

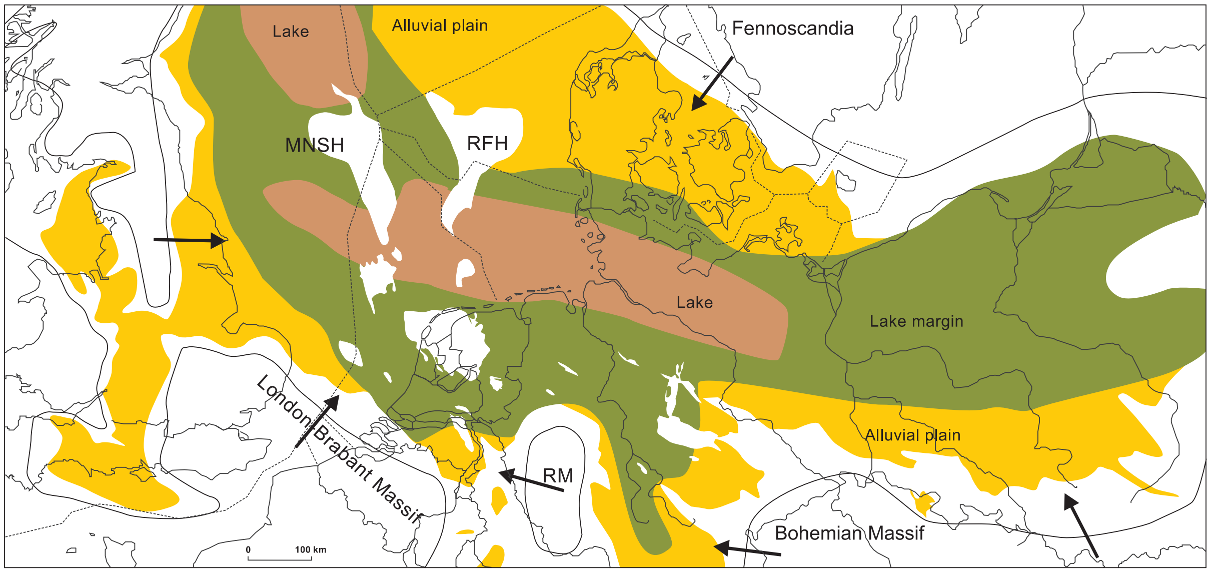


Fig. 1. Present-day distribution and facies map of the Lower Buntsandstein (Induan, Early Triassic). After Jubitz et al. (1987), Ziegler (1990), Cameron et al. (1992), Cope et al. (1992), Dadlez et al. (1998), Geluk (1999), Baldschuhn et al. (2001), Goldsmith et al. (2003) and Geluk (2005). Arrows indicate clastic influx. Solid lines represent the reconstructed basin outline. Abbreviations as in Fig. 8.

– the Upper Germanic Trias Group (Olenekian–Norian) comprising an alternation of fine-grained clastics, carbonates and evaporites with subordinate sandstones.

The boundary between these groups is formed by the Hardegsen or Base Solling Unconformity, which forms a regionally well-correlatable event (Ziegler, 1990; Geluk & Röhling, 1997, 1999; Geluk, 2005). The groups have distinct seismic characteristics and tectonic styles (Figs 5, 6). The top of the Upper Germanic Trias Group, i.e. the base of the Rhaetian Sleen Formation, forms an excellent marker on both seismic data and well logs. The Sleen Formation, representing the youngest Triassic, belongs lithos-

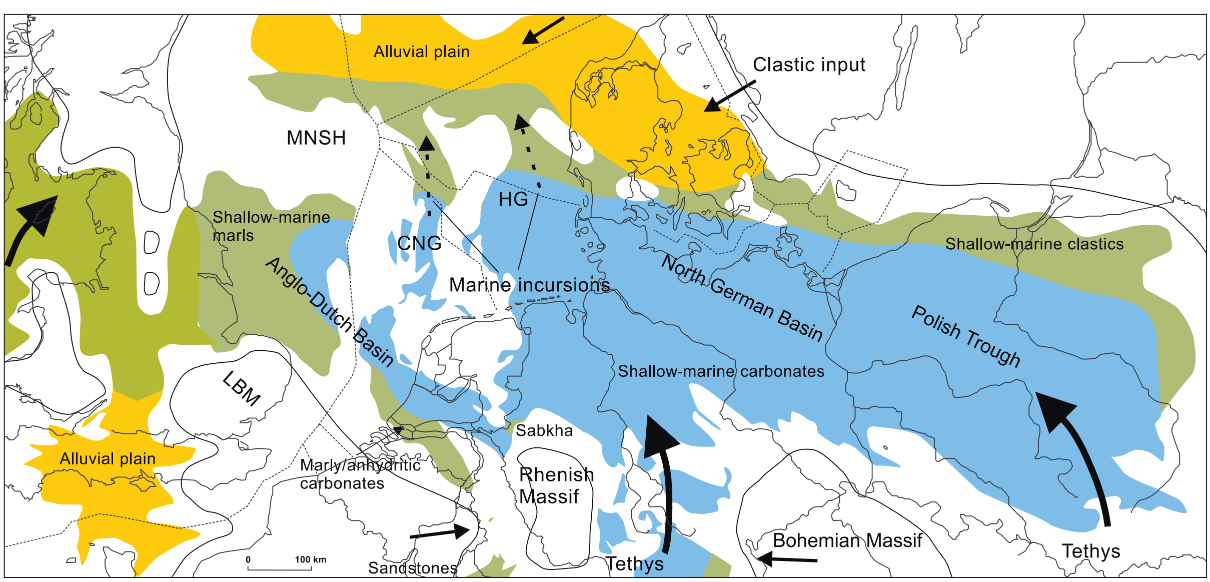


Fig. 2. Present-day distribution and facies map of the Lower Muschelkalk (Late Anisian, Middle Triassic). The thick arrows indicate the ingression from the Tethys through the Silesian-Moravian gate in the east and the more western Burgundy gate. After Jubitz et al. (1988), Cameron et al. (1992), Marek & Pajchlowa (1997), Geluk (1999, 2005) and Szulc (1999, 2000). Solid lines represent the reconstructed basin outline. Abbreviations as in Fig. 8.

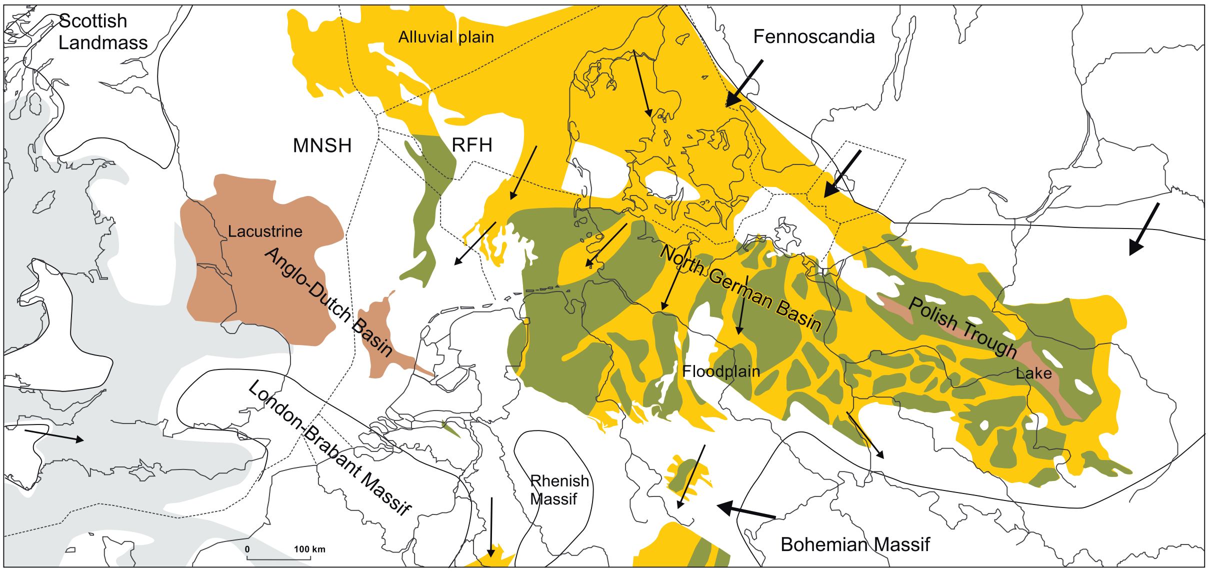


Fig. 3. Present-day distribution and facies map of the Schilfsandstein (Carnian, Late Triassic). The main sediment supply originated from Fennoscandia. After Beutler & Häusser (1981), Beutler, (1998), Jubitz et al. (1988), Cameron et al. (1992), Cope et al. (1992), Dadlez et al. (1998), Geluk (1999, 2005), Baldschuhn et al. (2001) and Goldsmith et al. (2003). Solid lines represent the reconstructed basin outline. Arrows indicate transport directions and clastic influx. The Schilfsandstein is also present in the area marked in grey in the west. Abbreviations as in Fig. 8.

tratigraphically to the Altena Group (Wong, this volume). For completeness, it is briefly discussed in this chapter.

Sequence-stratigraphically a distinction is made between the continental Induan–Olenekian and the marine-influenced younger Triassic. The Lower Buntsandstein Formation and Main Buntsandstein Subgroup (Induan–Olenekian) were controlled mainly by high-order Milankovich climatic cycles (Geluk & Röhling, 1999). In the Olenekian to Norian two transgressive–regressive mega-cycles are recognized. The first of these has its maximum flooding surface in the Late Ladinian Upper Muschelkalk and the maximum regression at the Ladinian–Carnian boundary. The second mega-cycle had its peak transgression in Late Carnian times, and the maximum regression in the Late Norian (Aigner & Bachmann, 1992; Aigner et al., 1999). These mega-cycles are recognized in a more or less similar form both in the Barents Sea and the Tethys (Gianolla & Jacquin, 1998).

Regional correlations

The stratigraphic framework of Table 1 can be tied very well to those adopted in the UK southern North Sea and Germany. This underlines the pronounced layer-cake character of the Triassic. Triassic units can be correlated over large distances throughout north-west Europe, as has been demonstrated by Geiger & Hopping (1968), Trusheim (1971), Beutler & Schüler (1987), Cameron et al. (1992), Geluk & Röhling (1997, 1999) and Geluk (2005). Correlations of the main epicontinental units in north-west Europe with the marine Triassic stages have been based mainly on palynology and paleobotany (Freudenthal, 1964; Visscher, 1966; Geiger & Hopping, 1968; Visscher & Commisaris, 1968).

The Lower Germanic Trias Group and its equivalents are a barren red-bed succession, and correlation relies entirely on lithological similarities and wireline-log patterns, whereas the underlying Zechstein Group and overlying Röt Formation can be successfully identified based upon their palynological assemblages. The age of the most marine succession in the Triassic, the Muschelkalk Formation, is proved by ceratites, whereas ostracods and palynomorph assemblages support the ages of the Keuper Formation.

Despite a great similarity in lithostratigraphic names, some differences occur in the definitions between the Netherlands and the surrounding countries. The base of the Triassic is picked significantly lower in the UK southern North Sea and Denmark than in the Netherlands and Germany. Rocks of the uppermost Zechstein in the Netherlands belong to the Triassic Bacton Group in the UK and Denmark. The top of the 'lithostratigraphic Triassic' is picked in the Netherlands and the UK at the base of the Sleen Formation or Penarth Group; in Germany at the top of the Mittelrhät. Other variations occur in the definitions of the Buntsandstein. The Main Buntsandstein Subgroup in the Netherlands comprises the Volpriehausen, Detfurth and Hardegsen formations (Van Adrichem Boogaert & Kouwe, 1994). In addition

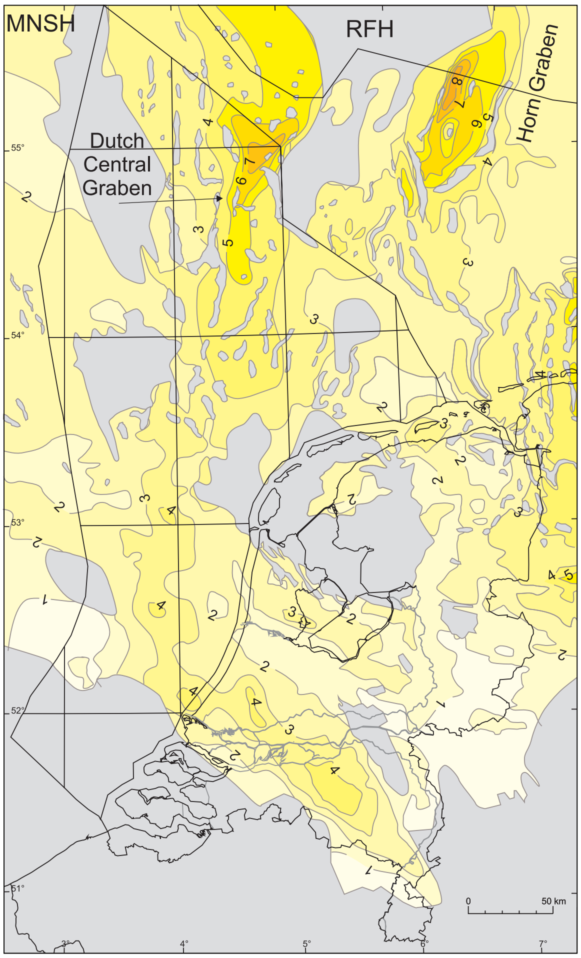


Fig. 4. Depth map (in kilometres) of the base of the Triassic. This map has been compiled from the geological atlas of the Netherlands (unpublished NITG compilation), with additional data from Day et al. (1980), Kockel (1995), Baldschuhn et al. (1996) and Johnson et al. (1999). Note the piercement by Zechstein salt in the northern half of the Netherlands. Abbreviations as in Fig. 8.

to these formations, the German Middle Buntsandstein Group includes the entire Solling Formation, whereas the British Bunter Sandstone Formation includes beds equivalent to the basal, sandy deposits of this formation.

Tectonic setting and events

Two extensional tectonic phases took place during the Triassic, related to the disintegration of Pangea (Wolburg, 1969; Ziegler, 1990; Röhling, 1991; Kockel, 1995):

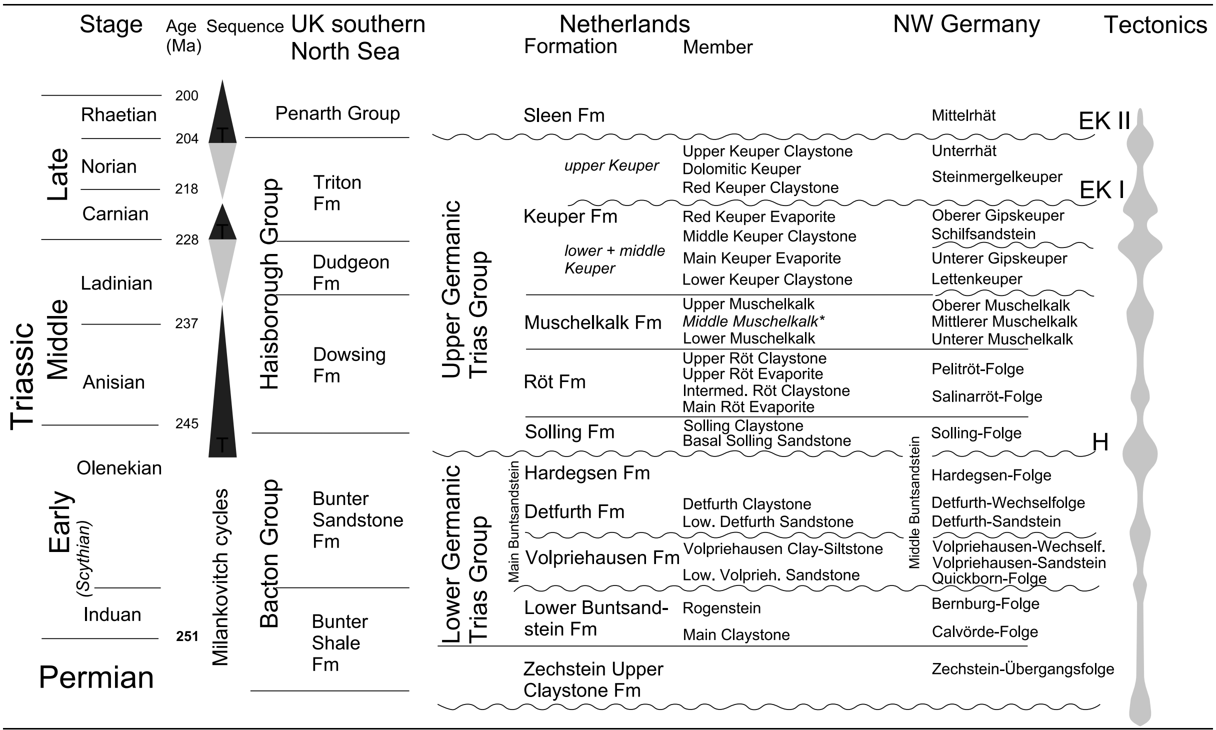
- the Hardegsen phase which culminated during the Olenekian; it affected the thickness and sand dispersal patterns of the Main Buntsandstein and comprises up to four short-lived rift pulses (Geluk & Röhling, 1997, 1999);
- the Early Kimmerian phase during Anisian to Norian times, which affected the thickness and salt distribution of the Röt, Muschelkalk and Keuper formations and which comprises up to five pulses (Wolburg, 1967; Beutler, 1995). It resulted in two unconformities; the most important lies at the base of the Red Keuper Claystone (Early Kimmerian I), the second at the base of the Sleen Formation (Early Kimmerian II).

These phases resulted in a progressive structural modification of the Southern Permian Basin, which became dissected by rifts, and split up into smaller units as can be seen in the isopach maps of the lithostratigraphic units discussed further on. In the Netherlands, relatively gentle faulting, uplift and erosion took place. The main rifts were the Glückstadt and Horn grabens in north-west Germany (Best et al., 1983; Geluk & Röhling, 1999). The differential subsidence decreased westward. The Main Buntsandstein reaches over 2500 m in the Glückstadt Graben and 3000 m in the Horn Graben, compared with 500 m in the Dutch Central Graben and 350 m in the Sole Pit Basin. A series of swells were formed, including the Netherlands Swell and the Cleaver Bank High (Fig. 7). On these swells, the Main Buntsandstein is less than 50 m thick (Geluk & Röhling, 1997, 1999). Uplift and subsequent erosion of these swells during the Jurassic Mid to Late Kimmerian phases removed much of their Triassic cover.

Analysis of the subsidence pattern shows that differential subsidence was spasmodic and shifted northwards with time, from the Roer Valley Graben and West Netherlands Basin to the Ems Low and the Dutch Central Graben (Geluk & Röhling, 1999). This was accompanied by strong uplift of the Mid North Sea and Ringkøbing-Fyn highs. The fault pattern of this phase suggests an east–west extension, with minor strike-slip movements along a number of fault zones.

The onset of the Early Kimmerian phase was in Anisian times, when differential movement started in the Dutch Central Graben. The Netherlands Swell was cut along a series of ESE–WNW trending faults, and partly collapsed. The Early Kimmerian movements continued intermittently during much of the Triassic, as is illustrated by the differential subsidence of the Dutch Central Graben and the movements along the Mid Netherlands Fault Zone. The strongest movements occurred during the Carnian. During these movements the main rift was situated in the Glückstadt Graben (Best et al., 1983; Frisch & Kockel, 1997). In the Netherlands, the Dutch Central Graben displayed the strongest subsidence. Basement faulting triggered widespread mobilization of Zechstein salt. Thick successions of the Muschelkalk and Keuper formations occur in the rim synclines of salt diapirs (Figs 5c, d). Ensuing strong uplift and subsequent deep erosion occurred in the southern Netherlands, where a system of NE-dipping fault blocks separated by WNW-trending faults was formed. The oldest deposits subcrop on the south-

Table 1. Stratigraphic subdivision of the Triassic in the Netherlands and adjacent countries.



After Van Adrichem Boogaert & Kouwe (1994), Johnson et al. (1994), Geluk (1999) and Kozur, 1999. Ages after ISC (2003); note that the only officially approved age is the Permian-Triassic boundary. Sequences after Gianolla & Jacquin (1998); transgressive sequences in black, regressive sequences in grey. EK I: main Early Kimmerian Unconformity, base Norian; EK II: Early Kimmerian II Unconformity, base Rhaetian; H: Hardegsen Unconformity. * The Middle Muschelkalk is an informal unit and comprises the Muschelkalk Evaporite and Middle Muschelkalk Marl.

western parts of these blocks, the youngest on the north-eastern parts (Figs 8, 9). Local intrusion of Zechstein salts occurred along the faults (NITG 1998, 2000). In the Dutch Central Graben these movements triggered further widespread mobilization of Zechstein salt and the collapse of salt pillows formed during the Early Triassic.

Apart from N–S trending faults, a system of WNW–ESE trending faults was active. These faults, for instance the Mid Netherlands Fault Zone and the North Dogger Fault Zone, are characterized by locally preserved Upper Triassic sediments in their hanging-wall blocks. Other faults (Gronau Fault Zone) mainly show uplift. They are interpreted here as transcurrent faults.

The Triassic displays a marked difference in tectonic styles in the Netherlands (Figs 5a–d, 6). In Figure 5a, the Triassic, affected by pre-Cretaceous erosion, is present as a unit of continuous reflectors with constant thickness. A small collapse graben, filled with Röt to Keuper sediments, developed here in response to the Early Kimmerian tectonics. These grabens are a few hundred metres up to 1 km wide. The Dutch Central Graben shows both symmetrical and asymmetrical Triassic depocentres, related to Zechstein salt withdrawal (Figs 5b–c). Figure 5b is a classical example of a turtle-back anticline, which was bordered by two growing salt pillows during the Triassic. In Figure 5c the salt structure to the west of this anticline moved into the diapiric stage during the Keuper, as indicated by the thickening of the succession towards the structure. Intrusion of Zechstein salt in the structure on the east in Figure 5c is of later date and related to the Late Cretaceous compressional tectonics. Triassic growth faults, often facilitated by Zechstein salt, are present at the western margin of the Dutch Central Graben (Fig. 5d), but also in the Ems Low and to the south-east of the Dutch Central Graben. Similar faults, interpreted as rift–raft zones, have been observed in Germany and the southern North Sea (Griffith et al., 1995; Best, 1996; Penge et al., 1999). The onset of the rifting shown in Figure 5d was during deposition of the Röt.

Lower Germanic Trias Group

The Lower Germanic Trias Group is divided into the Lower Buntsandstein, Volpriehausen, Detfurth and Hardegsen formations. The last three formations together form the Main Buntsandstein Subgroup (Table 1).

Lower Buntsandstein Formation

This formation is composed of a cyclic alternation of fine-grained lacustrine sandstones and clay-siltstones, which form stacked fining-upward sequences. These sequences,

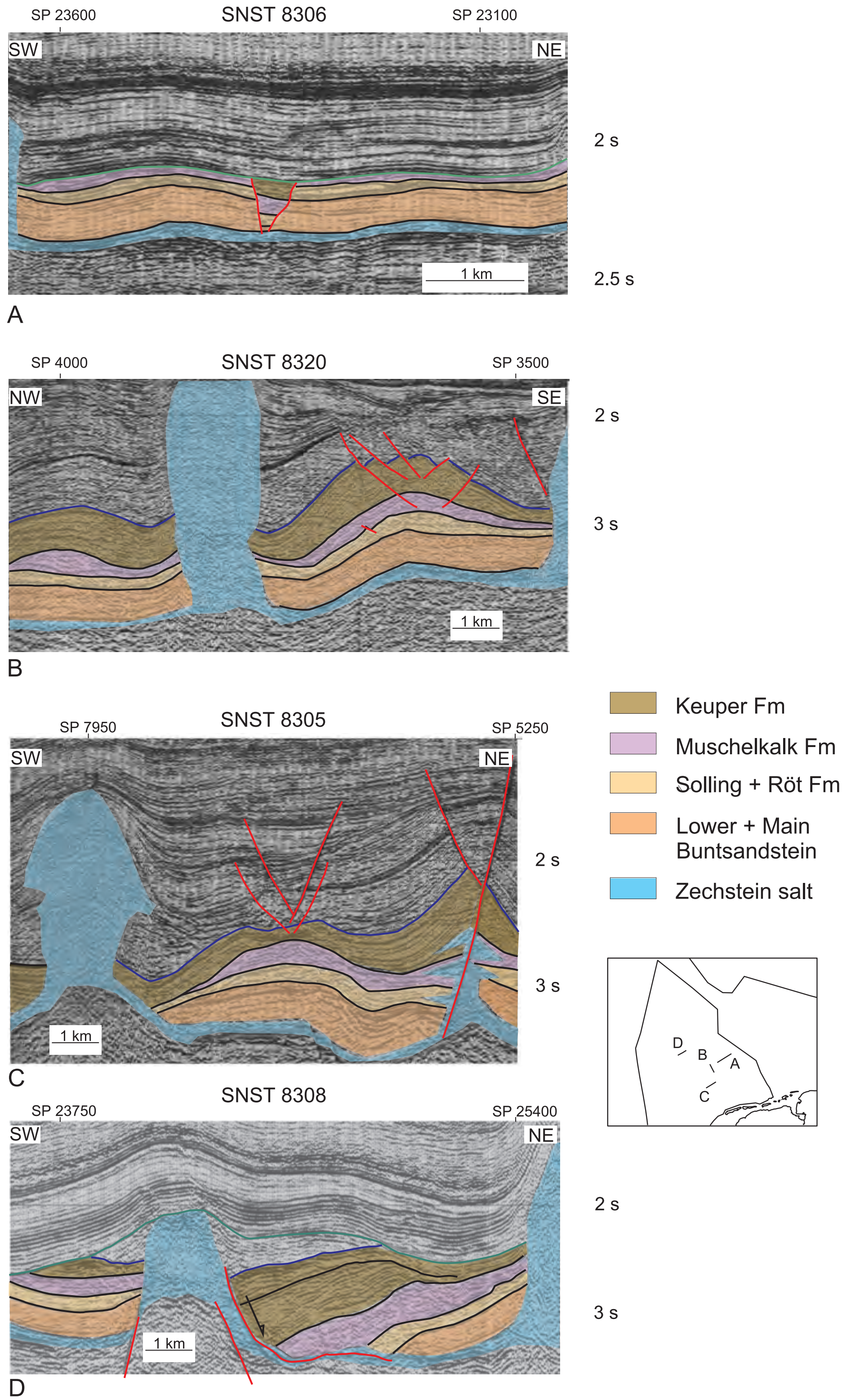


Fig. 5. Structural styles of the Triassic in the Netherlands. a) Seismic line SNST 8306. Narrow, 1-km-wide, graben embedded within a succession of Lower Buntsandstein to Muschelkalk, and filled with Röt, Muschelkalk and Keuper deposits, Schill Grund High. b) SNST 8320. Primary rim synclines with a thick Keuper succession, Dutch Central Graben. c) SNST 8305. Complex, faulted, Triassic depocentre, Dutch Central Graben. Intrusion of Zechstein salt is observed in the Röt, Muschelkalk and Keuper evaporite layers. d) SNST 8308. Rift–raft zone along a Triassic growth fault, western margin of Dutch Central Graben.

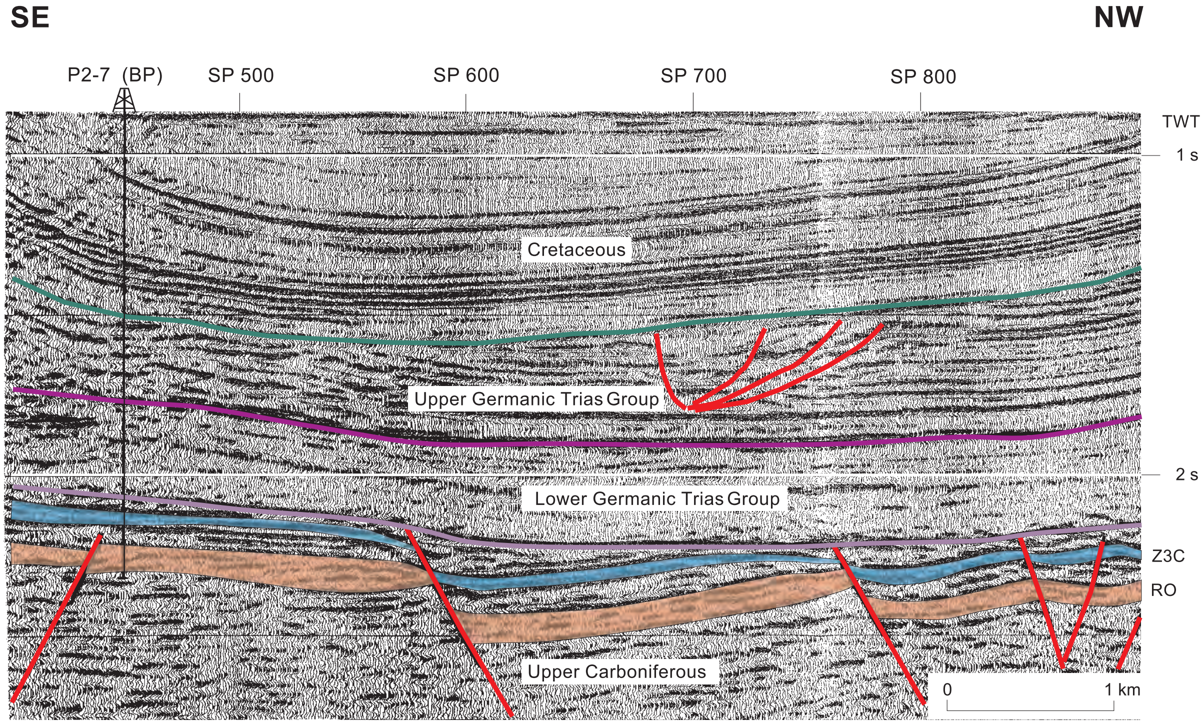


Fig. 6. Section from the Broad Fourteens Basin illustrating the seismic character of the Triassic. The Lower Germanic Trias Group is a rather transparent unit, with several continuous low-amplitude reflectors in its upper part. The Upper Germanic Trias Group displays an alternation of low and high-amplitude reflectors. The collapse graben in this group developed during the Late Kimmerian extensional tectonic phases. In this structure, the Muschelkalk Salt forms the detachment horizon. RO: Upper Rotliegend Group; Z3C: Z3 Carbonate.

20 to 40 m thick, can be correlated over distances of several 100 km (Geluk & Röhling, 1997, 1999). The development and thickness of the formation indicate a very uniform subsidence (Figs 9–11). The formation is of youngest Permian to Induan age (Kozur, 1999; Szurlies et al., 2003).

Strongest subsidence continued in the northern and western offshore areas, where the formation is up to 400 m thick. A topographically slightly more elevated area, the Netherlands Swell, was present in the north-central onshore area. At the eastern side of this swell, several, up to 5-m-thick limestone oolite beds are present (Rogenstein Member); elsewhere in the basin these beds are much thinner. The fine-grained character of the Lower Buntsandstein is also recorded north of the Mid North Sea High in the lower part of the British Smith Bank Formation (Goldsmith et al., 1995). This, combined with the absence of sandy deposits around this high, indicates that it was originally entirely covered by Lower Buntsandstein sediments.

At the southern margin of the basin, in the southern Netherlands and the adjacent German area (Wolburg, 1961), up to 200 m of massive sandstones are present. Conglomerates occur in the extreme south-east (Geluk, 1999).

Main Buntsandstein Subgroup

The Main Buntsandstein Subgroup displays a cyclic alternation of (sub-)arkosic sandstones and clayey siltstones organized in large-scale fining-upward sequences. Its deposition marks a fundamental reorganization of the subsidence and uplift patterns in response to the Hardegsen phase (Fig. 12). This phase comprises up to four individual pulses, the strongest of which occurred prior to the deposition of the Solling Formation. The formations of the subgroup, the Volpriehausen, Detfurth and Hardegsen formations, are tectono-stratigraphic units (Table 1; Fig. 11). Each consists of a first-order fining-upward cycle of basal sandstone and overlying clay-siltstones. Superimposed on this large-scale cycle is a hierarchical pattern of small-scale cycles (Geluk & Röhling, 1997, 1999).

In the southern Netherlands and the UK southern North Sea, the subgroup is entirely sandy (Cameron et al., 1992; Johnson et al., 1994; Geluk et al., 1996). The sandstones are mainly fluvial in the south and grade northward into predominantly eolian deposits (Fontaine et al., 1993; Ames & Farfan, 1996). The subsurface development in the Netherlands is similar to that in outcrops near Trier in the west of Germany (Mader, 1983). The clay-siltstones represent playa-lake deposits.

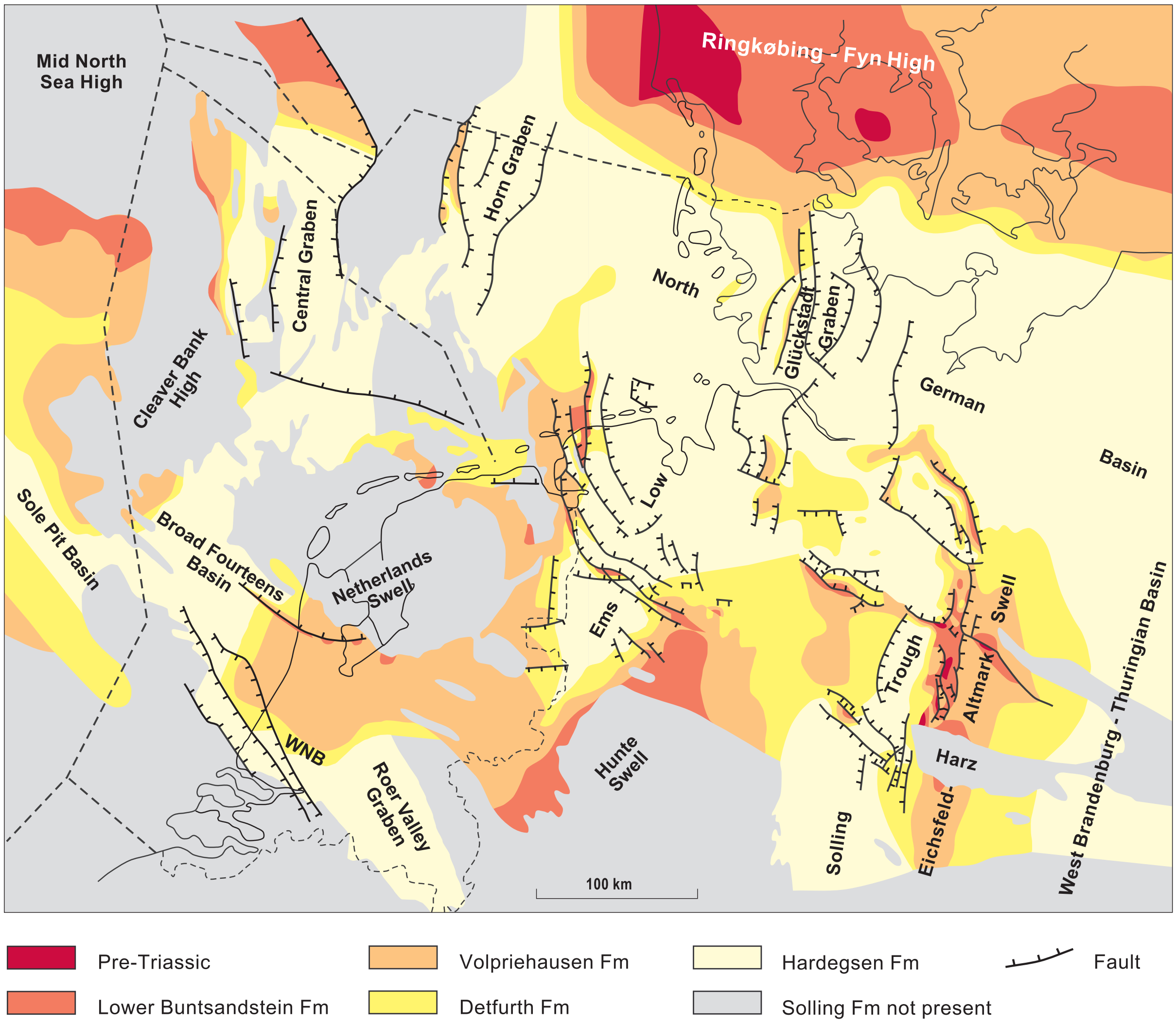


Fig. 7. Subcrop map of the Base Solling (= Hardegsen) Unconformity (after Geluk & Röhling, 1997, 1999). WNB: see Fig. 8.

Subsidence was strongest in the Dutch Central Graben, where locally over 500 m of sediments were deposited. The structure of this graben is complex as a result of extensive salt flow. In the Sole Pit Basin and western Netherlands offshore the thickness reaches 350 m, about as much as in the Roer Valley Graben and Ems Low.

Volpriehausen Formation

The Volpriehausen Formation displays its greatest thickness, over 200 m, in the Dutch Central Graben and the Broad Fourteens Basin. It reaches 100 m in the Ems Low and 150 m in the Roer Valley Graben. The Volpriehausen Unconformity at the base of the formation locally cuts up to several tens of metres into the Lower Buntsandstein Formation.

The *Lower Volpriehausen Sandstone* displays marked differences in thickness (Fig. 13). The sandstone is arkosic, with a quartz content slightly below 50%. It is cemented by high percentages of calcite and dolomite, especially in its lower part. In the Dutch Central Graben, salt plugging of the pores in the sandstone is common (Fontaine et al., 1993; Dronkert & Remmelts, 1996; Purvis & Okkerman, 1996). In the Ems Low the member reaches a thickness of over 20 m, thinning towards the Netherlands Swell to less than 5 m. West of this swell its thickness rapidly increases to up to 100 m in the Broad Fourteens Basin. This can be explained in part by a westward facies transition from clay-siltstones into sandstones, and by the presence of an older sandstone unit which is absent on the swell (Geluk & Röhling, 1999). Two depocentres were situated in the Dutch Central Graben, with over 60 m of sandstones deposited. North of the Dutch offshore, the sands thin to 10 m. In southern areas, sands were deposited in a fluvial setting, whereas in the northern areas, predominantly eolian sands occur (Fontaine et al., 1993; Ames & Farfan, 1996).

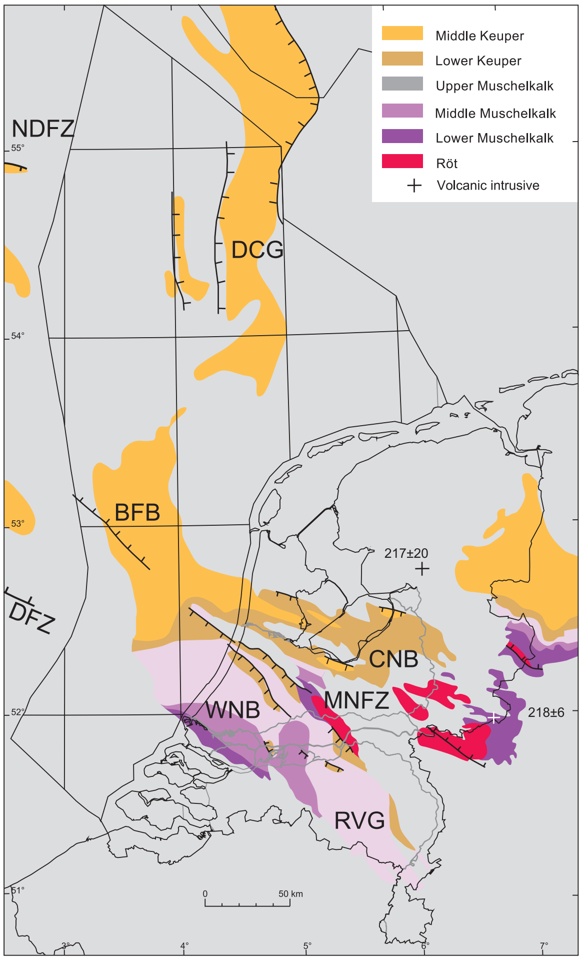


Fig. 8. Subcrop map of the Early Kimmerian Unconformity (base Red Keuper Claystone or, where this unit is not present, Sleen Formation). UK part based on Cameron et al. (1992), German part on Wolburg (1967, 1969), Hilden (1988) and Baldschuhn et al. (2001). Contemporaneous volcanic intrusives (ages shown) have been encountered in two wells (Sissingh, 2004). After Geluk (1999). Abbreviations, including those used in other figures: BFB: Broad Fourteens Basin; CBH: Cleaverbank High; CNB: Central Netherlands Basin; CNG: Central North Sea Graben; DFZ: Dowsing Fault Zone; DCG: Dutch Central Graben; EL: Ems Low; HG: Horn Graben; LBM: London-Brabant Massif; MNFZ: Mid Netherlands Fault Zone; MNSH: Mid North Sea High; NDFZ: North Dogger Fault Zone; RFH: Ringkøbing-Fyn High; RM: Rhenish Massif; RVG: Roer Valley Graben; WNB: West Netherlands Basin.

The *Volpriehausen Clay-Siltstone* forms a succession of predominantly lacustrine siltstones and marls, with subordinate sandstones. A number of carbonate oolite beds occur. In the southern offshore of the Netherlands the siltstones grade into fluvial and eolian sandstones (Ames & Farfan, 1996). The sandstones are more fine-grained than the Lower Volpriehausen Sandstone and cemented by dolomite, calcite and ankerite (Geluk et al., 1996). The member displays considerable variations in thickness, which have been attributed to erosion prior to deposition of the Detfurth Formation.

Detfurth Formation

The occurrence of the Detfurth Formation is restricted to the Early Triassic lows as a result of uplift and erosion prior to deposition of the Solling Formation. The depositional thickness of the formation displays considerable variation: 60–100 m in the Dutch Central Graben, 50–80 m in the Ems Low and 20–40 m in the West Netherlands Basin, Roer Valley Graben and Broad Fourteens Basin. In the West Netherlands Basin and Roer Valley Graben the formation consists entirely of sandstones. At its base the Detfurth Unconformity cuts into the Volpriehausen Formation. In the southern and western offshore areas of the Netherlands this unconformity is the most prominent one in the Buntsandstein (Geluk et al., 1996; Geluk & Röhling, 1999).

The isopach map of the *Lower Detfurth Sandstone* indicates a further reorganization of the subsidence pattern by the third rift pulse of the Hardegsen phase (Fig. 14). The sandstone has a relatively high quartz-grain content (>50%) and is loosely quartz-cemented. In the basin-margin area, the sandstone is one of the best hydrocarbon reservoirs in the Main Buntsandstein Subgroup. In the Dutch Central Graben, its porosity is often destroyed by salt plugging (Fontaine et al., 1993; Dronkert & Remmelts, 1996; Purvis & Okkerman, 1996).

Previous areas of strong subsidence during deposition of the Volpriehausen Formation, e.g. the Broad Fourteens Basin, became inactive, and depocentres migrated north and eastwards to the Ems Low and the Dutch Central Graben, where up to 60 m of sandstones accumulated. The different lows were clearly separated from each other. On the flanks of the Cleaverbank High and the Netherlands Swell, the Lower Detfurth Sandstone is locally absent due to non-deposition, or to erosion before the deposition of the Detfurth Claystone (Geluk & Röhling, 1999).

The *Detfurth Claystone* follows the thickness trend of the Lower Detfurth Sandstone. It thins towards the Netherlands Swell and Cleaver Bank High, but probably once covered both swells. It is composed of claystones, with thin intercalations of siltstone. In the southern offshore and onshore, the claystones grade laterally into eolian sandstones (Ames & Farfan, 1996).

Hardegsen Formation

The Hardegsen Formation consists predominantly of siltstones, with subordinate, thin sandstone beds. Significant amounts of sandstone occur only in the basin-margin area. The present-day thickness of the formation was strongly influenced by the pre-Solling erosion and displays strong variations. Only erosional remnants remain; in the

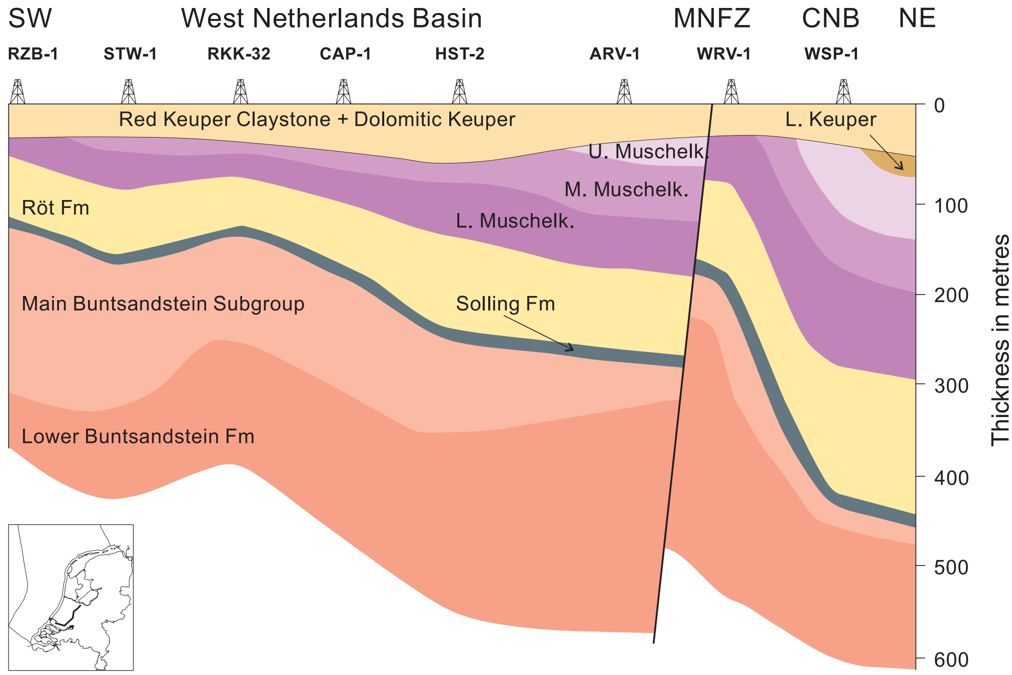


Fig. 9. Section based on well data, showing the structure of the Triassic in the West Netherlands Basin. Note the Early Kimmerian Unconformity at the base of the Red Keuper Claystone (upper Keuper). The base of the Sleen Formation serves as the reference level for this section. After Geluk (1999, 2005). CNB, MNFZ: see Fig. 8.

Ems Low and in the Dutch Central Graben (well F9-3) up to 200 m occur, while in the Horn Graben (Danish well S-1) over 500 m have been encountered. In the Broad Fourteens Basin, the West Netherlands Basin and the Roer Valley Graben, the thickness reaches up to 70 m. During deposition of the formation, syn-rift subsidence occurred in the Dutch Central Graben, as is evident from the thickening of individual sequences. In other earlier depocentres, such as the Broad Fourteens Basin, the West Netherlands Basin and the Roer Valley Graben, differential subsidence had ceased (Geluk & Röhling, 1999).

Upper Germanic Trias Group

The Upper Germanic Trias Group comprises the Solling, Röt, Muschelkalk and Keuper formations (Table 1; Fig. 15). The Base Solling Unconformity forms the base of the group, its top is the Rhaetian transgression surface at the base of the Sleen Formation. The original distribution of this group before the erosion related to the Mid and Late Kimmerian tectonic phases was much greater than that of the remnants presently preserved in the main Late Jurassic basins (Geluk, 2005).

Solling Formation

The Solling Formation rests on various older deposits, from which it is separated by the Base Solling or Hardegsen Unconformity (Fig. 7). It was deposited after the fourth pulse of the Hardegsen phase and comprises a basal sandstone, overlain by fine-grained deposits. The base of the formation becomes progressively younger to the west, accompanied by a marked decrease in thickness from 125 to 10 m (Fig. 16). The main subsidence occurred in the Dutch Central Graben and the Ems Low. A stable platform area with up to 25 m of Solling deposits occupied the major part of the western offshore and southern onshore. Only the youngest part of the formation is present here.

The *Basal Solling Sandstone* occurs in the northern half of the Netherlands, but only locally exceeds 10 m in thickness (Fig. 16). In north-west Germany the equivalent sandstones are up to 60 m thick and related to major fault trends (Beutler et al., 1992). In the eastern Netherlands, the position of the sandstones relative to salt structures points to a relation with salt flow. Eolian sands filled the depressions created by salt migration (NITG, 2000).

The *Solling Claystone* is a succession of siltstones and claystones, more than 100 m thick, in the Ems Low and the Horn and Dutch Central grabens. Within this member, two sequences can be distinguished, separated by an unconformity (Geluk & Röhling, 1999). The lower part has a variable thickness, and is limited to the northern half of the Netherlands. The upper part has a more sheet-like character and originally probably covered the entire Dutch area. In the southern Dutch Central Graben, an intra-Solling sandstone of eolian origin is present, up to 100 m thick; it forms the main producing unit in the L9-

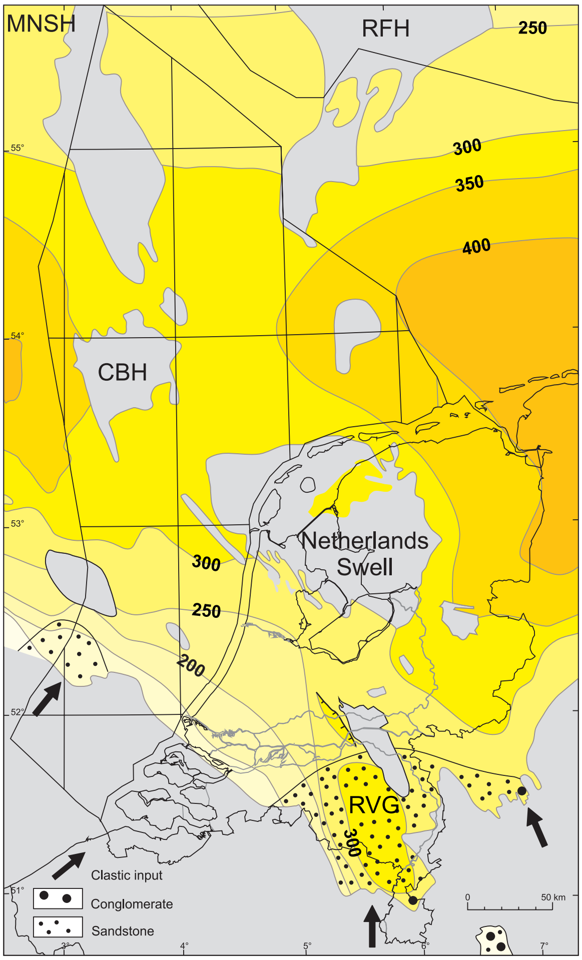


Fig. 10. Isopach map (in metres) of the Lower Buntsandstein Formation. Data in Belgium after Demyttenaere (1989), UK part after Cameron et al. (1992) and Johnson et al. (1999), Germany after Kockel (1995), Baldschuhn et al. (1996, 2001) and Boigk (1961). Abbreviations as in Fig. 8. After Geluk (1999, 2005).

FF field (De Jager & Geluk, this volume; Geluk, 2005). The sandstone is present in a wedge-like depression similar to the one shown on Figure 5d. A number of wells in the vicinity of active fault zones encountered similar, thinner, sandstones (Fig. 16). For the sandstone the name Middle Solling Sandstone Member has been proposed (Geluk, 2005).

Röt Formation

The Röt Formation comprises a lower evaporitic part, and an upper part dominated by clay and siltstones. Strongest subsidence occurred in the Dutch Central Graben, where over 300 m of Röt Formation accumulated, and in the Ems Low and Roer Valley Graben. In the south-western offshore the formation is less than 50 m thick, and consists only of its youngest part (Fig. 17).

The *Main Röt Evaporite* has a wide distribution. Halite was deposited in a wide area, including the former Netherlands Swell and Cleaverbank High, both of which are areas with a reduced Main Buntsandstein succession (Figs 12, 17). The thickness of the halite varies from 30 to over 150 m in the Broad Fourteens and Central Netherlands basins. In the Dutch Central Graben, up to 300 m of halite occur. The southernmost limit of the Röt halite runs parallel to the margin of the Central Netherlands Basin, suggesting some degree of fault control on the salt deposition. South of this limit, the member comprises anhydrites.

The *Intermediate Röt Claystone* is made up of anhydrite-bearing clay and siltstones. The *Upper Röt Evaporite* is halite-bearing only in the Dutch Central Graben, the eastern Netherlands and adjacent parts of Germany; its thickness is up to 20 m.

The *Upper Röt Claystone* comprises clay and siltstones with intercalated carbonates in its upper part. In the southern onshore this member grades partially into sandstones of the *Röt Fringe Sandstone* (Fig. 17). These record the last important clastic influx from southern source areas into the Netherlands during the Triassic (Geluk et al., 1996).

Muschelkalk Formation

The Muschelkalk Formation constitutes a distinctly marine interval. It consists of lower and upper carbonate parts, and a middle evaporitic part. The carbonate parts comprise a cyclic alternation of limestones and marls, which can be correlated over hundreds of kilometres (Gaertner & Röhling, 1993). During deposition, the main subsidence, up to 500 m, was in the southern Dutch Central Graben. Low subsidence rates occurred on the flanks of the London-Brabant Massif and the Mid North Sea and Ringkøbing-Fyn highs. The formation is up to 250 m thick in the Broad Fourteens Basin, and was much affected by Early Kimmerian uplift and erosion in the central and eastern Netherlands (Figs 9, 18).

The *Lower Muschelkalk* is composed of a cyclic alternation of limestones, dolomites and marls. Along the London-Brabant Massif and the Mid North Sea High, an anhydritic, marly carbonate succession less than 50 m thick was deposited, whereas in the Dutch Central Graben and the Ems Low the member expands to over 150 m. In the eastern Netherlands, deposition took place in a very shallow sea with fluctuating salinities (NITG, 1998). The nearest sand deposition took place in the Mechernich and Trier areas in western Germany (Hilden, 1988).

The Lower Muschelkalk forms the oldest outcropping rocks in the eastern Netherlands; before 1902, however, these were regarded as Wealden (Lower Cretaceous). Once their Middle Triassic age was established (Müller, 1902), the area became interesting for coal and salt exploration. The rocks are quarried east of Winterswijk, where a cyclic

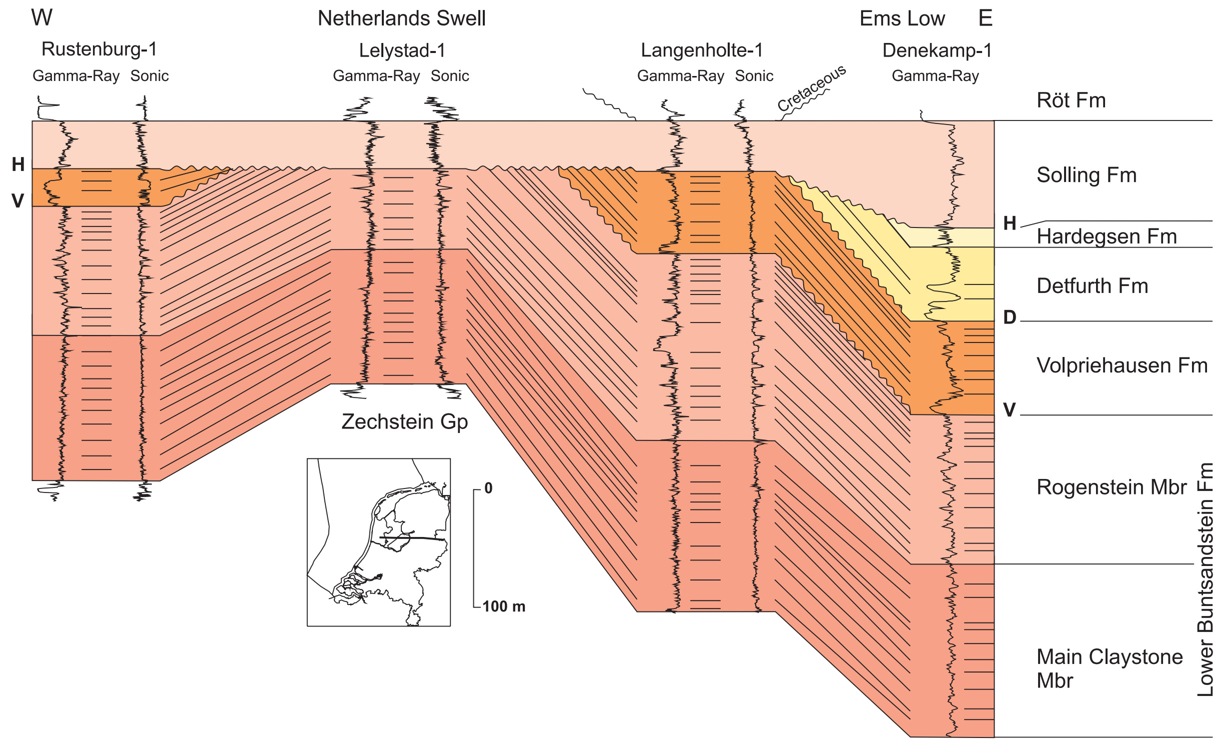


Fig. 11. Log correlation of the Lower Germanic Trias Group. The Solling Formation overlies this group unconformably. The high-resolution sequences in the group are well-correlatable. On the Netherlands Swell, erosion occurred into the Rogenstein Member during the Hardegsen tectonic phase (after Geluk & Röhling, 1997, 1999). Unconformities: V: Volpriehausen, D: Detfurth, H: Hardegsen. See Table 1 for stratigraphic nomenclature.

succession of shallow-marine marls and dolomites is exposed (Harsveldt, 1973). The dolomitization caused strata-bound vuggy porosity. A similar porosity contributes to the gas production from this member in the De Wijk field (Pipping et al., 2001). Cyclically developed mud-crack horizons occur. The quarry is famous as a vertebrate track-site, and has yielded a rich fossil and mineral collection (Oosterink, 1986; Diedrich, 2001; Oosterink et al., 2003).

The *Middle Muschelkalk* comprises the Muschelkalk Evaporite and the Middle Muschelkalk Marl members (Fig. 15). The Muschelkalk Evaporite contains halite in many areas (Fig. 18); outside these areas anhydrite is present. In the south-west part of the Netherlands, dolomites are found locally (Borkhataria, 2004). The occurrences of halite in the hanging-wall blocks along the Mid Netherlands Fault Zone are remarkable from a regional point of view and indicate fault movements during evaporite deposition. This is in line with similar occurrences in north-west Germany, where up to six halite cycles, separated by dolomitic claystone beds have been identified in this member (Brückner-Röhling, 1999). The complete succession only occurs in fault-bounded depressions; the total halite thickness exceeds 500 m (Wolburg, 1967; Gaertner & Röhling, 1993). The Netherlands appears to have occupied a more marginal position: only two to three cycles are recognized, with a total thickness of up to 50 m. In the Dutch Central Graben and the Ems Low, over 100 m of halite occur locally. The Middle Muschelkalk Marl, a marly dolomite of 20 to 50 m thick, covers the Muschelkalk Evaporite.

The *Upper Muschelkalk* comprises an alternation of carbonate and claystone beds. Its thickness is typically around 50 m but reaches over 125 m in the Dutch Central Graben and the Ems Low. Westward the carbonate content gradually decreases, but important carbonate intervals known from Germany, the Trochitenkalk and Ceratitenkalk (cf. Gaertner & Röhling, 1993), can be identified throughout the Netherlands.

Keuper Formation

The base of the Keuper Formation is marked by claystones with a low sonic velocity. Deposition of these beds started earlier in the western than in the eastern Netherlands. The formation was strongly affected by the Early Kimmerian rifting and related salt movement (Frisch & Kockel, 1997). Within the formation a number of unconformities reflects this intermittent tectonic activity (Wolburg, 1967, 1969; Beutler & Schüler, 1978, 1987; Schröder, 1982). Figures 19 and 20 show the isopach maps of the Keuper below and above the Early Kimmerian Unconformity respec-

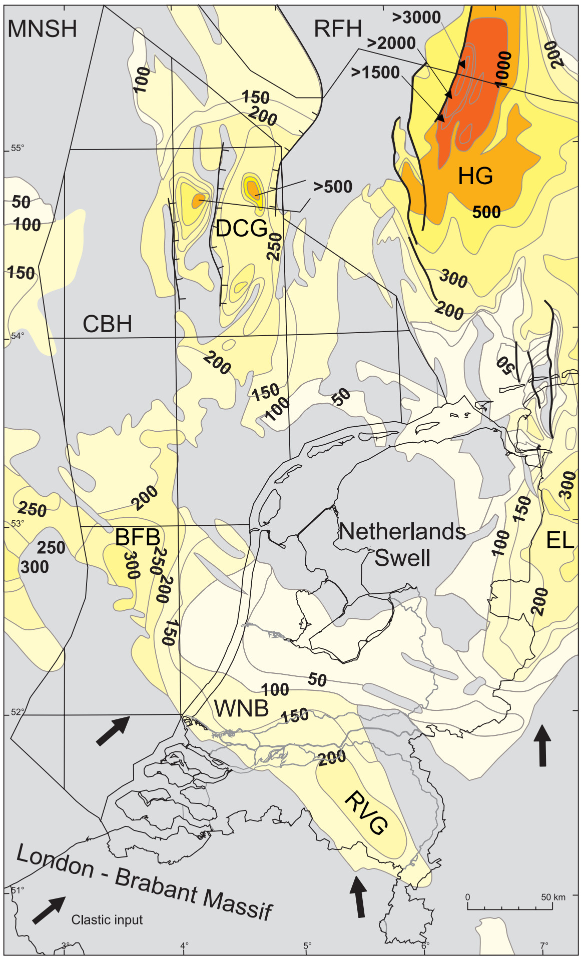


Fig. 12. Isopach map (in metres) of the Main Buntsandstein Subgroup. UK part after Cameron et al. (1992), Germany after Boigk (1961), Kockel (1995) and Baldschuhn et al. (1996). In the Danish and German North Sea sectors, the subgroup reaches a thickness of over 3000 m in the Horn Graben. Abbreviations as in Fig. 8. After Geluk (1999).

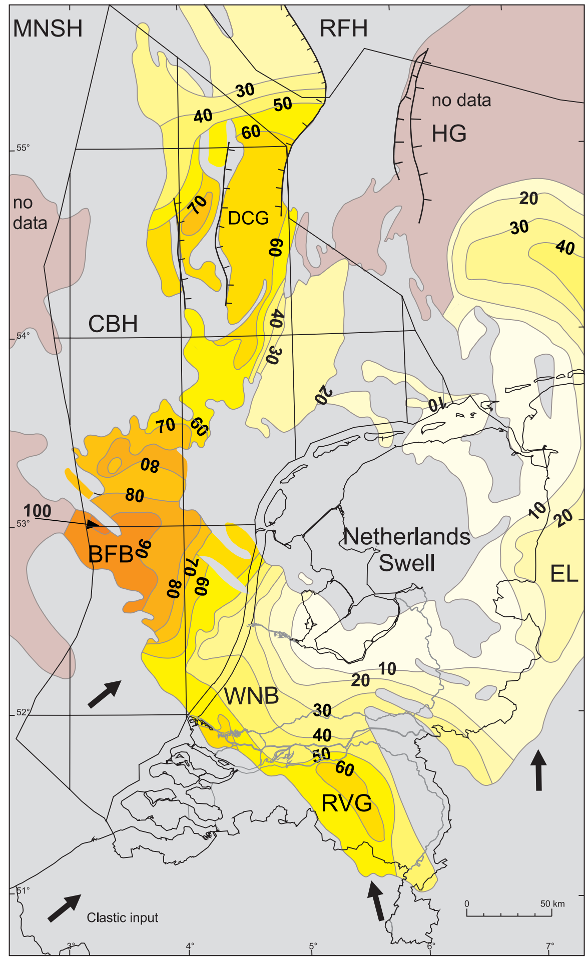


Fig. 13. Isopach map (in metres) of the Lower Volpriehausen Sandstone. The German part has been compiled from Boigk (1961), Baldschuhn et al. (1996) and Röhling (1999). Note the great thickness in the Broad Fourteens Basin; this is explained in the text. Abbreviations as in Fig. 8. After Geluk (1999).

tively. The pre-unconformity succession is referred to as the lower and middle Keuper, the post-unconformity succession as the upper Keuper. This subdivision is informal, and differs from the definition of Lower, Middle and Upper Keuper as used in Germany (Table 1).

The thickness of the lower and middle Keuper displays a great variation (Fig. 19). Areas of strong differential subsidence were situated in the Dutch Central Graben and the Ems Low, with over 1000 m, and in the northern P quadrant, with over 400 m. In adjacent parts of Germany, the thickness reaches locally 1000 m in the Ems Low and 5000 m in the Glückstadt Graben (Best et al., 1983; Frisch & Kockel, 1997). Undrilled halfgrabens, with thick Keuper sediment wedges, occur in the Central Graben (Fig. 5d). In the southern Netherlands, most of the lower and middle Keuper was already removed during the Early Kimmerian phase; in other areas this occurred during the Mid to Late Kimmerian phases. Isolated occurrences were encountered in the hanging-wall blocks along the Mid Netherlands, the North Dogger and several other fault zones.

The *Lower Keuper Claystone* comprises reddish and dark-coloured claystones, alternating with thin layers of dolomite, fine-grained sandstone and coal. Towards the top anhydrite beds appear. The greatest thickness, up to 200 m, occurs in the southern part of the Dutch Central Graben. In other areas, the thickness of the member is typically between 40 and 80 m.

The *Main Keuper Evaporite* comprises anhydrites, halite and claystones. It attains its greatest thickness in the Dutch Central Graben (> 400 m). In graben structures in the Ems Low, the thickness increases locally to more than 1000 m (Frisch & Kockel, 1997). The member is halite-bearing in the Dutch Central Graben, the Ems Low, and locally in the Broad Fourteens Basin.

The *Middle Keuper Claystone* is mainly composed of

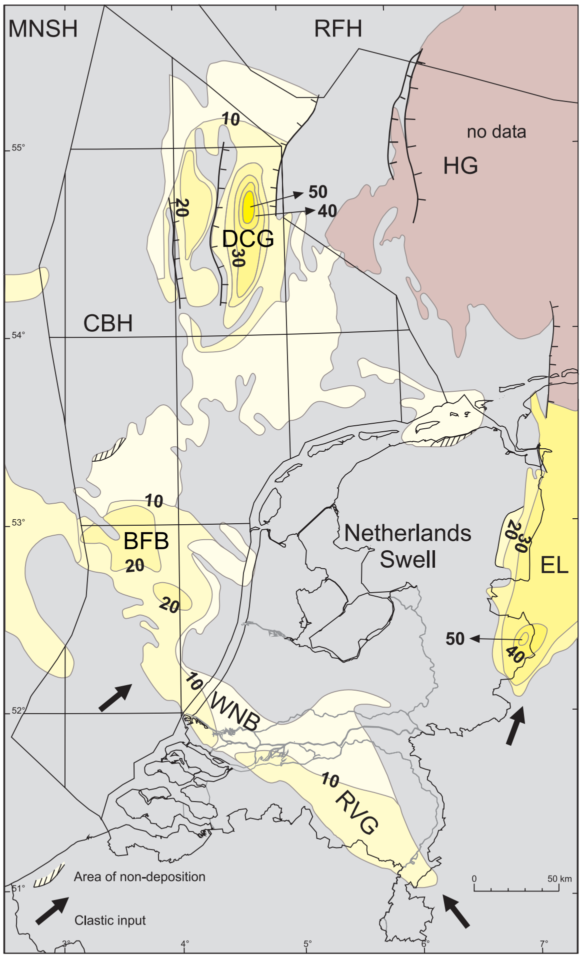


Fig. 14. Isopach map (in metres) of the Lower Detfurth Sandstone. Note the two small areas where no sand was deposited. Abbreviations as in Fig. 8. After Geluk & Röhling (1999).

reddish claystones. Locally in the southern Dutch Central Graben, however, it includes in its basal part up to 10-m-thick, mineralogically immature, mica-bearing, grey to green-coloured, high-gamma-ray sandstones that are the equivalents of the German *Schilfsandstein* (Beutler & Schüler, 1978). The member reaches a thickness of 100 m in the Dutch Central Graben and in the P quadrant; in other areas it varies between 10 and 50 m. In northwest Germany, the Schilfsandstein rests unconformably upon older Keuper deposits (Wolburg, 1969; Beutler, 1995; Frisch & Kockel, 1997, 1999). In the Netherlands, it overlies the Lower Gipskeuper with one exception in the southern Netherlands where, based on palynological evidence, Middle Keuper Claystone has been found overlying uppermost Muschelkalk (well Nederweert-1; NITG, 2001).

The *Red Keuper Evaporite* consists largely of anhydrite; halite only occurs in the member in the Dutch Central Graben and in Germany (Wolburg, 1967; Trusheim, 1971). The thickness of the member ranges from 2 to 40 m in the southern areas and reaches 200 m in the southern Dutch Central Graben.

The *Red Keuper Claystone* is separated by the Early Kimmerian I Unconformity from the underlying Triassic (Figs 8, 9). Together with the Dolomitic Keuper and the Upper Keuper Claystone members it forms the post-rift succession of the Early Kimmerian phase. The strongest differential subsidence during deposition of the upper Keuper occurred in the Dutch Central Graben (> 200 m) and the Broad Fourteens Basin (> 100 m). Fault movement is also evident along the Mid Netherlands Fault Zone. The upper Keuper covered most of the paleorelief, except in the east , where Rhaetian deposits overstep onto the unconformity (Fig. 20). The Red Keuper Claystone comprises variegated clay and siltstones. The *Dolomitic Keuper* contains light-coloured micritic carbonates and, in the West Netherlands Basin, anhydrite layers. The *Upper Keuper Claystone* is made up of dark-coloured marls and claystones.

Sleen Formation

This formation, of Rhaetian age and belonging to the Altena Group, represents the youngest part of the Triassic (Table 1). It rests directly above the Early Kimmerian II Unconformity on older Triassic rocks. The formation comprises a basal part of grey, fossiliferous, marine claystones, overlain by brown, locally sandy claystones (often containing a considerable quantity of megaspores). Generally, the Sleen Formation is easily recognizable on wire-line logs by its uniform thickness, relatively high gamma-ray and low acoustic-velocity readings. The base of the formation forms a prominent seismic marker. The thickness of the formation reaches nearly 70 m in the Dutch Central Graben.

Paleogeography

Despite the patchy distribution of mainly the youngest Triassic formations shown on the isopach maps, the entire area north of the London-Brabant Massif formed during the Triassic part of a large sedimentary basin (Geluk, 2005). Locally, this sedimentation was interrupted by uplift or erosion as a result of the Hardegsen and Early Kimmerian phases.

The deposition of the Triassic was governed initially by the inherited Permian basin configuration, at a paleolatitude around 20° N. During Late Permian times the connection with the Barents Sea was disrupted and an increased influx of fresh water and clastics, originating from the south, gradually transformed the Zechstein evaporite basin into a large shallow playa lake (Fig. 1). Mainly fine-grained sediments were deposited, except in the southern part of the Netherlands where coarser clastics were laid down in the Roer Valley Graben by ephemeral braided rivers. To a lesser extent, sands were redeposited by north-

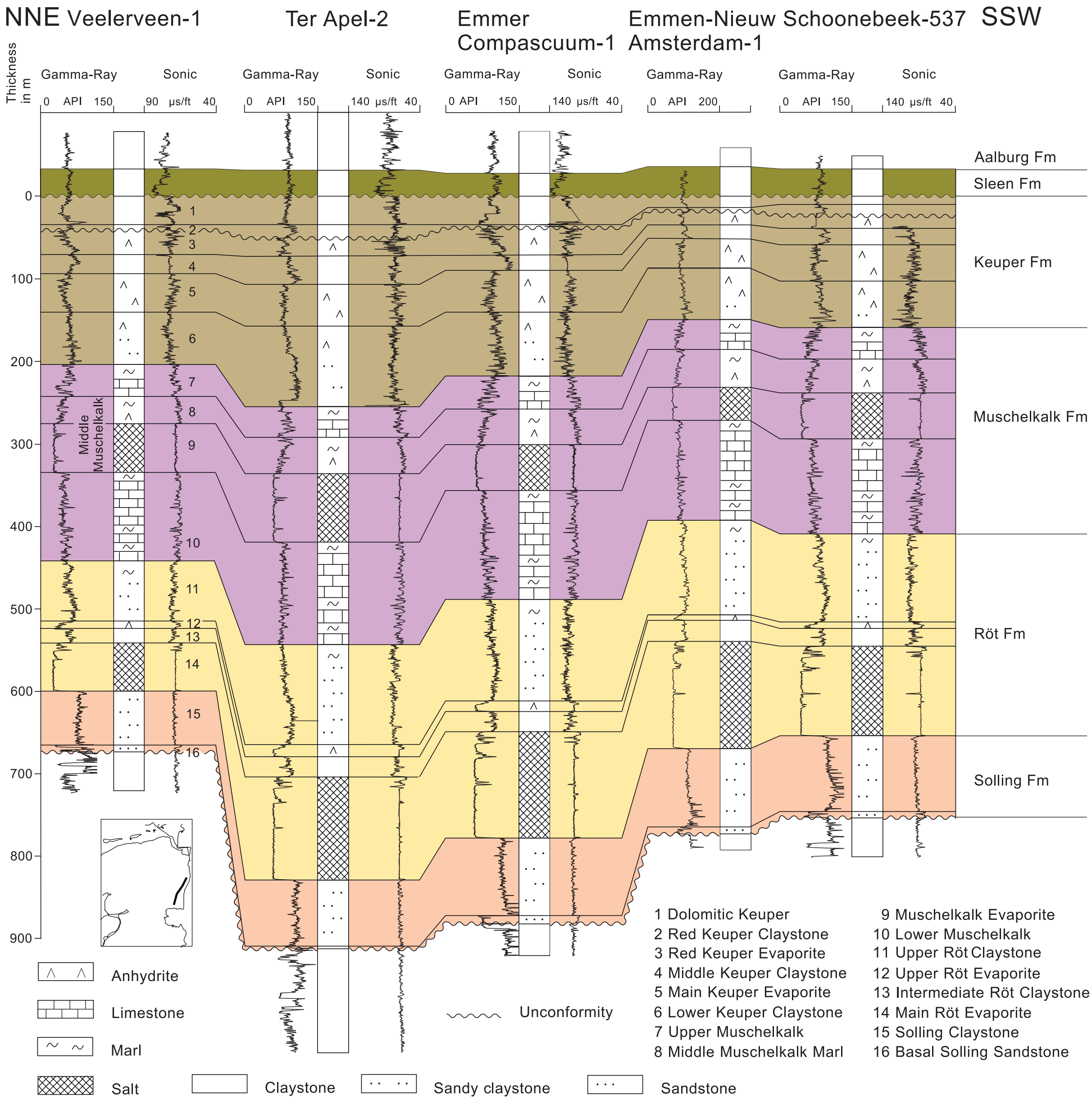


Fig. 15. Log correlation of the Upper Germanic Trias Group, eastern Netherlands. This profile illustrates the layer-cake character of this group over large areas. Reference level is the base of the Sleen Formation. After NITG (2000).

eastern trade winds as dunes. The deposition of the Lower Buntsandstein was governed by climatic Milankovich cycles of various magnitudes. These cycles caused basin-wide base-level variations (Geluk & Röhling, 1997, 1999). A succession of laterally persistent, fining-upwards cycles was laid down. In the central Netherlands a subtle, mostly submerged, swell developed, around and on top of which carbonate oolite beds developed. This swell corresponds with parts of the Permian Texel-IJsselmeer High and older, Carboniferous structures. Similar to the model proposed by Voigt & Gaupp (2000) for the Thuringian Basin in Germany, it is envisaged that the oolites formed during periods of fair weather. During storms, redeposition of oolites occurred over large areas around the swells. Some marine influence occurred in the eastern, Polish, part of the basin (Beutler & Szulc, 1999; Roman, 2004).

A major fluvial system in the south-eastern Netherlands supplied clastics from the Armorican Massif, the Massif Central and the Vosges Mountains during the Early to Middle Triassic (Ziegler, 1990). Additional clastics originated from the London-Brabant Massif (Geluk et al., 1996). During the Induan, these clastics were effectively trapped in the Roer Valley Graben. During the Olenekian, uplift of the hinterland in combination with

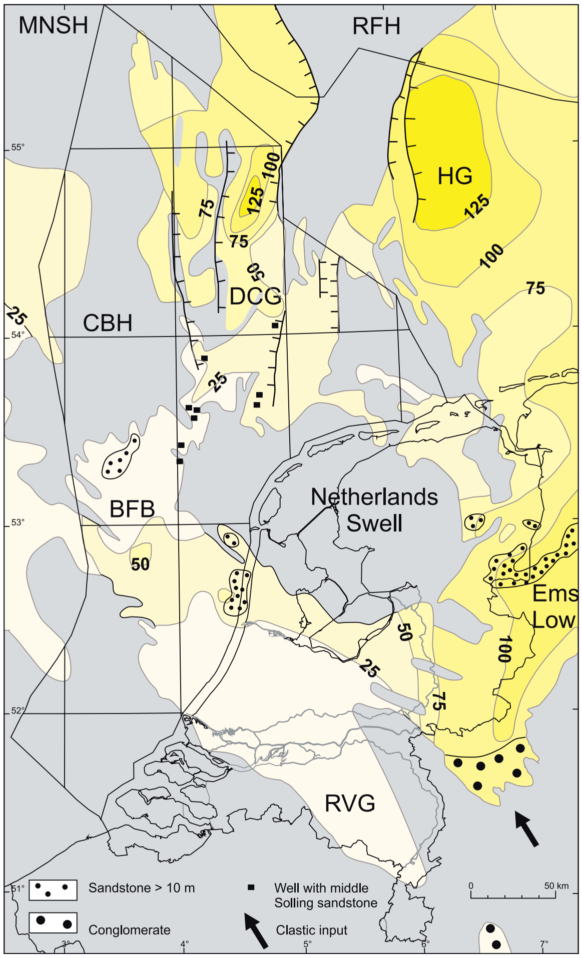


Fig. 16. Isopach map (in metres) of the Solling Formation. UK part based on Cameron et al. (1992), German part after Boigk (1961), Beutler et al. (1992) and Baldschuhn et al. (1996). Abbreviations as in Fig. 8. After Geluk (1999).

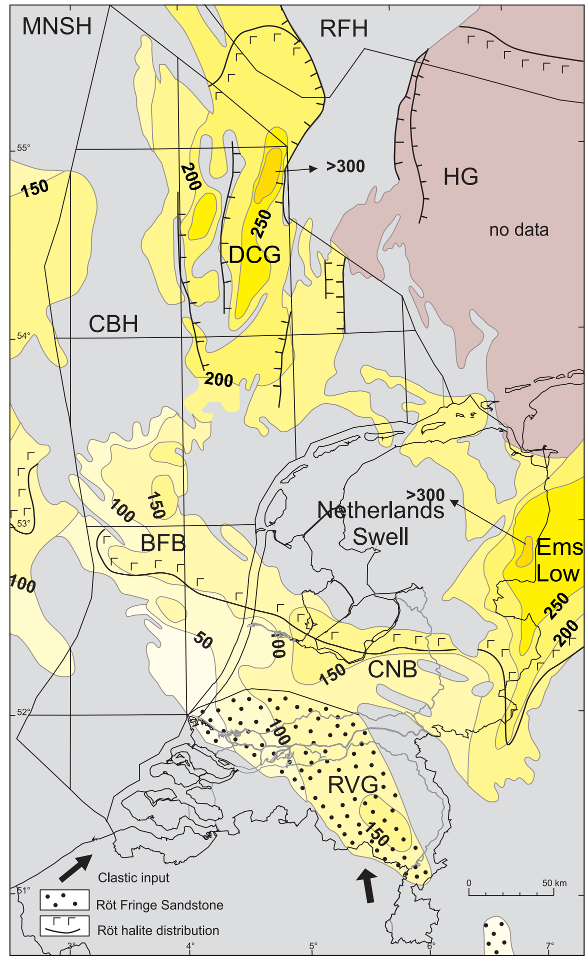


Fig. 17. Isopach map (in metres) of the Röt Formation. The map also shows the limits of salt distribution and the area of the Röt Fringe Sandstone in the south. Abbreviations as in Fig. 8. After Geluk (1999).

the Hardegsen rift phase resulted in an increase of clastic material transported into the basin. In response to humid periods in the hinterland, fluvial systems built out northwards through the Roer Valley Graben and West Netherlands Basin (Fig. 21); a second branch was situated more to the east in the Ems Low. In central parts of the basin, redeposition of fluvial sands into dune fields occurred on a wide scale during dry periods. During low clastic influx, the playa lake expanded again towards the margins of the basin. The repetition of these processes caused the cyclic alternation in the Main Buntsandstein Subgroup. The larger cycles, represented, for example, by the Volpriehausen and Detfurth formations, are tectonically driven; the high-resolution sequences within these formations are climate-driven (Van der Zwan & Spaak, 1992; Geluk & Röhling, 1997).

Several pulses of tectonic extension dissected the former Southern Permian Basin into smaller elements. The accommodation space in these rifts changed with time, shifting during the Olenekian deposition of the Main Buntsandstein Subgroup from the Roer Valley Graben into the Dutch Central Graben and the Ems Low (Figs 13, 14; Geluk & Röhling, 1999). Thickening of the subgroup towards the London-Brabant Massif (Fig. 12) suggests the presence of a growth fault at the southern margin of the West Netherlands Basin. The Lower Volpriehausen Sandstone and the Lower Detfurth Sandstone are fluvial to eolian in the southern areas (Ames & Farfan, 1996) and the Ems Low, and mainly eolian in the Dutch Central Graben (Fontaine et al., 1993). Extensional tectonics during deposition of the Main Buntsandstein Subgroup resulted in rapid subsidence of the Dutch Central, Horn and Glückstadt grabens, and contemporaneous uplift of a number of NNE–SSW trending swells in the Netherlands. On these swells, erosion removed much of the initially deposited Main Buntsandstein cover (Figs 7, 11).

On top of the Hardegsen Unconformity, fluvio-lacu-

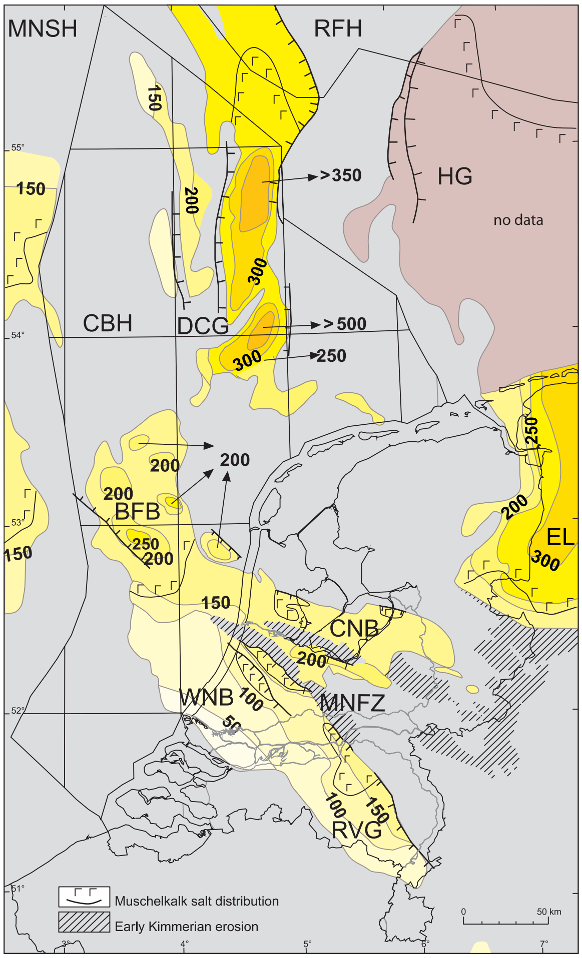


Fig. 18. Isopach map (in metres) of the Muschelkalk Formation, also showing the distribution of Muschelkalk salt. German part compiled from Wolburg (1967), Schröder (1982), Hilden (1988) and Gaertner & Röhling (1993). UK part based upon Cameron et al. (1992). After Geluk (1999) and NITG (2001, 2002). Abbreviations as in Fig. 8.

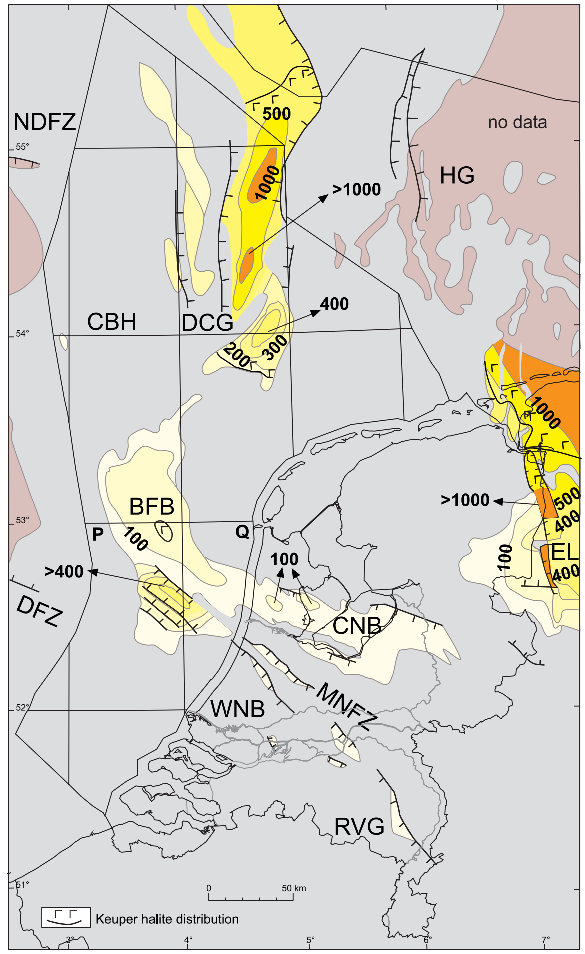


Fig. 19. Isopach map (in metres) of the lower and middle Keuper, also showing the limits of Keuper salt. UK part based on Cameron et al. (1992) and Griffiths et al. (1995). German part compiled from Wolburg (1967), Schröder (1982), Kockel (1995), Baldschuhn et al. (1996) and Frisch & Kockel (1997). After Geluk (1999) and NITG (2001, 2002). Abbreviations as in Fig. 8.

strine deposits of the Solling Formation covered the relief. The Solling is considered a post-rift deposit, which covered all structural elements in a sheet-like way. In part, however, the Solling was deposited contemporaneously with rifting as is expressed by sand-filled, fault-bounded grabens.

In Early Anisian times, a connection was established with the Tethys Ocean via the Silesian-Moravian Gateway, allowing marine influences to enter the basin once more. This resulted in a transgressive mega-sequence, which prograded westwards (Ziegler, 1990). Temporary interruption of this connection with the ocean resulted in the deposition of the Röt Evaporite. The remainder of the Röt Formation was deposited in a vast, shallow, brackish lagoon, which received the last influx of clastics from southern sources (Geluk et al., 1996). The formation was deposited under an extensional tectonic regime, with the Dutch Central Graben as the main area of differential subsidence. A number of faults, e.g. the Mid Netherlands Fault Zone, became active, dissecting previous Early Triassic structures (e.g. Netherlands Swell).

The Muschelkalk Formation was deposited during the maximum transgression, when coarse clastic deposition was pushed back to the margins of the basin and the sea covered most intra-basinal highs (Fig. 2). In the Netherlands, being situated in the western parts of the large basin, the Muschelkalk carbonates are not as well developed as in Germany. Their deposition took place in a shallow, epeiric sea. As a result of Late Anisian tectonic movements, the connection with the Tethys was interrupted and evaporites were deposited. A number of grabens formed within the basin and thick salts accumulated; these were the Dutch Central Graben, the Westdorf

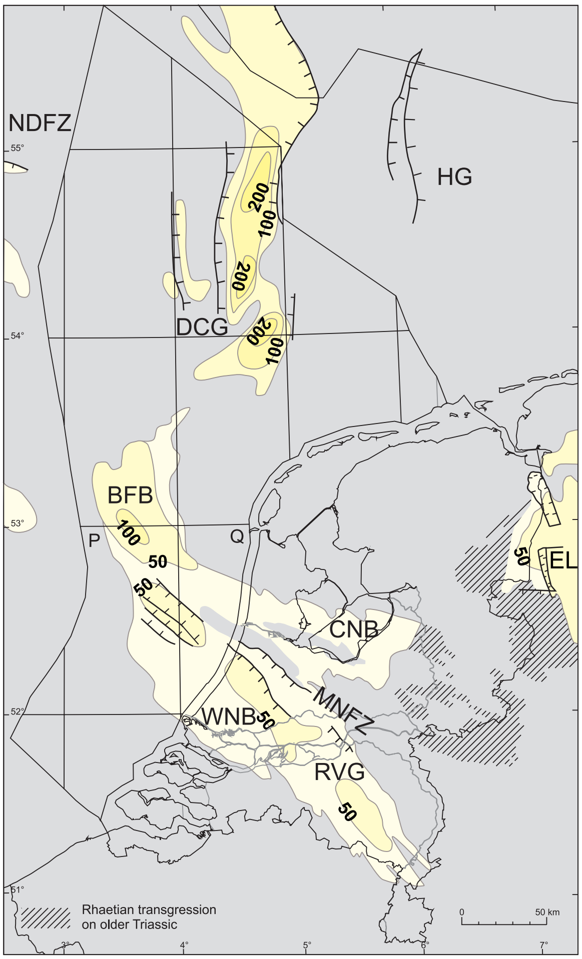


Fig. 20. Isopach map (in metres) of the upper Keuper. The upper Keuper is absent below the Early Kimmerian Unconformity in the indicated area of the Rhaetian transgression. UK part based on Cameron et al. (1992), German part adapted from Wolburg (1967). Abbreviations as in Figure 8. After Geluk (1999) and NITG (2001, 2002).

and Glückstadt grabens in Germany, and the West Netherlands and Central Netherlands basins (Geluk, 1999; Geluk et al., 2000; Baldschuhn et al., 2001).

Uplift of the Fennoscandian Shield during the Early Ladinian occurred in response to the Early Cimmerian orogeny, marking the closure of the paleo-Tethys (Ziegler, 1988). It resulted in the supply of clastic material to the basin from the north-east (Bertelsen, 1980; Beutler & Schüler, 1987; Ziegler, 1990). This renewed clastic supply ended the deposition of carbonates, and marks the base of the Keuper Formation, which prograded eastwards with time (Szulc, 2000). During the deposition of the Keuper, the Netherlands received mainly fine-grained sediments, transported over a great distance from the Fennoscandian Shield (Fig. 3). Deposition took place in brackish to normal-marine, shallow conditions. In Carnian times, Early Kimmerian, extensional tectonics caused rapid subsidence of the grabens where thick successions of evaporites were deposited. Rifting had shifted almost entirely into the Dutch Central Graben, the Ems Low and local fault zones elsewhere. In the graben, basement fault movement triggered widespread diapirism of Zechstein salt, as witnessed by thick successions of the Keuper in rim synclines. Some volcanic intrusions accompanied these tectonics (Fig. 20; Sissingh, 2004). Contemporaneous uplift of the London-Brabant Massif and surrounding areas resulted in widespread erosion. After these movements ceased, sedimentation was resumed during the Norian, and the Red Keuper Claystone and Dolomitic Keuper, together equivalent to the German Steinmergelkeuper, were deposited unconformably upon older Triassic rocks. These units were not deposited in the eastern Netherlands, where the Rhaetian Sleen Formation covered the remaining relief (Fig. 20).

Economic geology

Triassic rocks are important in the Netherlands for the following resources:
- natural gas, mainly occurring in the Lower Germanic Trias Group;
- rock salt from the Röt Formation, exploited by means of solution-mining;
- limestone and dolomite from the Lower Muschelkalk, quarried in the eastern Netherlands.

Natural gas

Triassic rocks form the second largest gas reservoir in the Netherlands, after the Permian Upper Rotliegend. The total initial gas reserve amounts to 299×10^9 m^3. The reservoirs contain also oil and condensate. The gas is sourced mainly from the Upper Carboniferous coal measures, and partly also from the basal Namurian hot shales (Fontaine et al., 1993; De Jager et al., 1996; De Jager & Geluk, this volume). The gas accumulations in the Triassic occur in areas with no or insufficient sealing of the Upper Rotliegend by Zechstein salts. Reservoirs are mainly formed by the Main Buntsandstein and the Röt Formation in the south-western onshore (De Jager et al., 1996). The Solling Claystone seals the accumulations in the Main Buntsandstein (Spain & Conrad, 1997), the clayey Muschelkalk and Keuper succession those in the Röt. Oolite beds of the Lower Buntsandstein and dolomite beds in the Muschelkalk form local reservoirs in the eastern Netherlands (Gdula, 1983; Bruijn, 1996; Pipping et al., 2001).

Oil

Lower Triassic reservoirs in the south-western Netherlands contain substantial volumes of oil. The oil was sourced by the juxtaposed Jurassic Posidonia Shale Formation, and possibly also by Carboniferous rocks (De Jager et al., 1996).

Rock salt

Rock salt (halite, NaCl) is produced from the Röt Formation near Hengelo in the eastern Netherlands. Production started in 1919 in the Buurse concession, and was transferred northwards to the Twenthe-Rijn concession during the early 1930s. The salt is produced by solution mining from depths between 300 and 500 m (Harsveldt, 1980; NITG, 1998). The annual production in 2000 amounted to 1881 kt (Geluk et al., this volume).

Limestone and dolomite

Limestone and dolomite of the Lower Muschelkalk have been quarried east of Winterswijk in the eastern Netherlands since 1933. The material is mainly used in road-building and agriculture (Van der Meulen et al., this volume).

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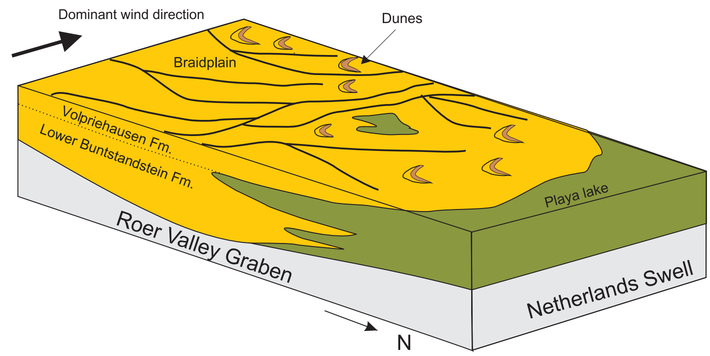

Fig. 21. Depositional model for the Lower Buntstandstein and Volpriehausen formations in the southern Netherlands. Yellowish colour indicates predominantly sandstones, greenish colour predominantly siltstones. During deposition of the Lower Buntsandstein, the fluvial systems were confined to the Roer Valley Graben; during deposition of the Volpriehausen Formation they built out to the north.

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Jurassic

Th.E. Wong

ABSTRACT

The Dutch Jurassic, here described together with the Rhaetian and Ryazanian, comprises four lithostratigraphic groups. The lower group is laterally uniform, mainly argillaceous and marine; it has a Rhaetian to Oxfordian age and represents essentially a pre-rift sequence. Its thickness reaches more than 1800 m. The three other groups, one mainly marine, one largely continental, and one continental to restricted marine, are generally coarser siliciclastic sediments. They represent more or less contemporaneous Callovian to Ryazanian syn-rift deposits in different basins and are locally more than 2500 m thick. In addition, a localized volcanic formation exists. During the Triassic and Jurassic the Netherlands changed structurally from being part of the larger Southern Permian Basin into several smaller, fault-bounded basins and highs. This change involved two extensional phases in Jurassic times: Mid Kimmerian (Aalenian-Callovian/Oxfordian) and Late Kimmerian (Kimmeridgian-Ryazanian). The basins can be grouped into three structural provinces: the N-S striking northern province (Dutch Central Graben), the E-W striking eastern province (Lower Saxony Basin) and the NW-SE striking central and southern system (e.g. Roer Valley Graben-Broad Fourteens Basin). In the Rhaetian and Early Jurassic, the area underwent a relatively constant subsidence. Later on, structural intricacy augmented gradually, especially during the Mid-Kimmerian phase. Various Jurassic sandstones are reservoirs for oil and gas. The Toarcian Posidonia Shale Formation sourced several relatively small oil accumulations.

Keywords: Rhaetian, Ryazanian, Kimmerian rifting, Netherlands, stratigraphy, basin development

Introduction

Previous research

Various publications deal specifically with the regional stratigraphy and geological setting of the Jurassic in the southern North Sea area (Brown, 1990; Michelsen & Wong, 1991; Cameron et al., 1992; Underhill & Partington, 1993; Partington et al., 1993). An integrated regional interpretation of the Jurassic in north-west Europe was presented by Ziegler (1990). Publications by Ten Dam & Reinhold (1942), Faber (1946), Harsveldt (1963), Haanstra (1963), Heybroek (1974), Van Wijhe (1987) and Burgers & Mulder (1991) deal exclusively with the Jurassic geological history in the context of the Dutch setting. The increasing availability of exploration data spurred the publication of several papers on specific Jurassic basins: Herngreen et al. (1984, 1991, 2000), Herngreen & Wong (1989), Wong et al. (1989) and Dronkers & Mrozek (1991). This culminated in the revision of the then existing stratigraphy by working groups consisting of members from the Geological Survey of the Netherlands (RGD) and NOGEPA, the organization of oil and gas-producing companies in the Netherlands (Van Adrichem Boogaert & Kouwe, 1993-1997). In that compilation, a sequence-stratigraphic subdivision for the Upper Jurassic has been introduced. Recently, Herngreen et al. (2003) wrote a comprehensive overview of the Jurassic geology of the Netherlands. They presented a stratigraphic framework showing the relationship between the north-west European standard ammonite zonation and selected biostratigraphic datum levels based on dinoflagellates, sporomorphs, foraminifers and ostracods. They also included a detailed correlation with the sea-level curve of Haq et al. (1988), and a scheme showing correlations with Germany, Denmark and the UK (Fig. 1). Like the overview by Herngreen et al. (2003), the present chapter uses the conventional North European or Boreal subdivision of the Late Jurassic into Oxfordian, Kimmeridgian and Portlandian. Moreover, to avoid a mixed Boreal (Portlandian) and Tethyan or Mediterranean (Berriasian) nomenclature, the name Ryazanian is applied for the earliest Cretaceous stage (cf. Herngreen & Wong, 1989, this volume). Abbink et al. (2004) demonstrated the importance of 'high-resolution' stratigraphy for the Jurassic-Lower Cretaceous deposits of the southern part of the Dutch Central Graben. Based on the quantitative distribution pattern of terrestrial palynomorphs they recognized various sea-level fluctuations and climate changes in that area. The Netherlands Institute of Applied Geoscience TNO included a chapter on the Jurassic paleogeographic and structural development in their atlas dealing with the onshore subsurface geology (NITG, 2004). Based on TNO data, Duin et al. (2006) recently published a series of thickness and depth maps, some of which have been used in this chapter.

Geological setting

During the Triassic and Jurassic the structural outline of the Netherlands progressively changed from one single, extensive basin, the Southern Permian Basin, into

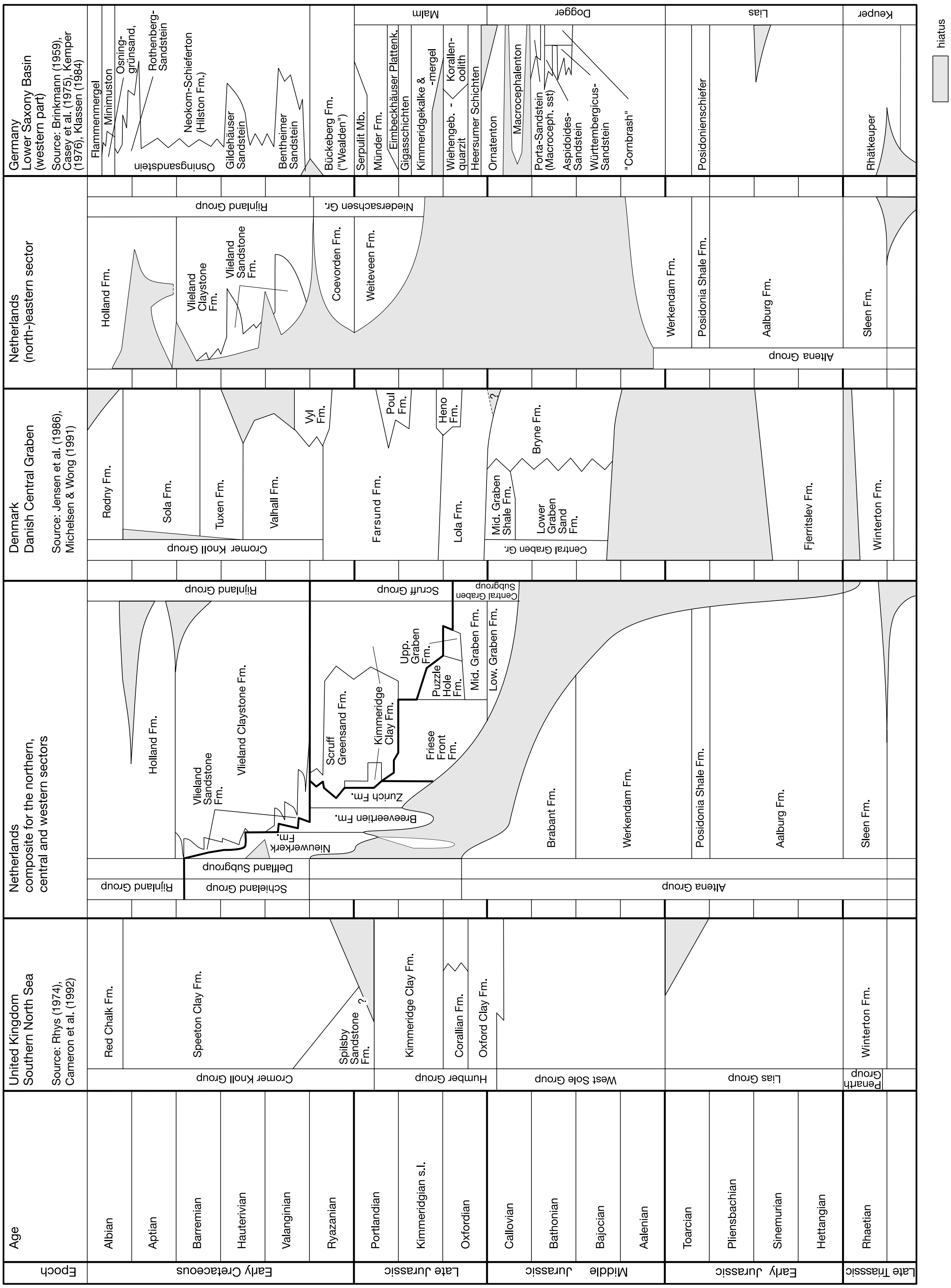


Fig. 1. Regional stratigraphic correlation chart of the Jurassic and Lower Cretaceous for the Netherlands and nearby areas (after Van Adrichem Boogaert & Kouwe, 1993-1997; Herngreen et al., 2003).

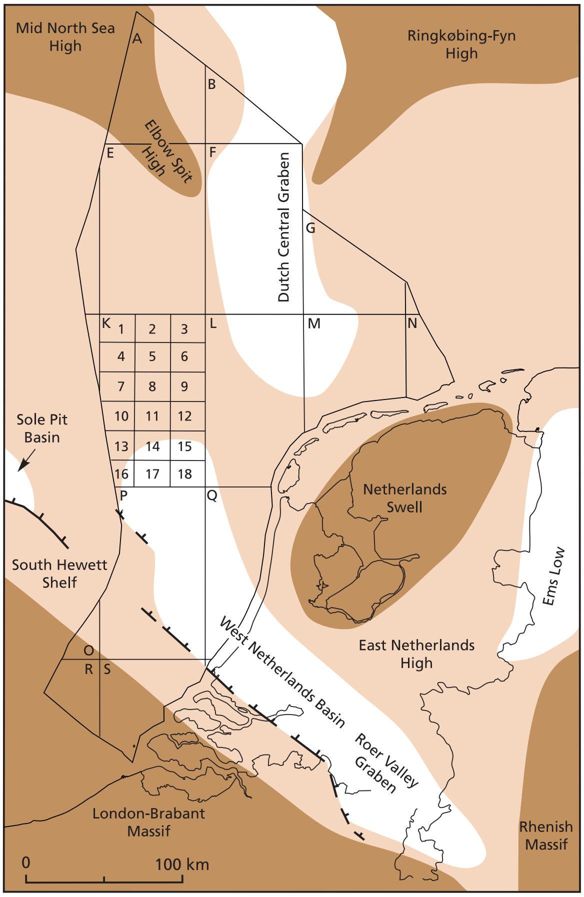


Fig. 2. Map of latest Triassic to Middle Jurassic structural elements in the Netherlands (after Van Adrichem Boogaert & Kouwe, 1993-1997). Dark brown: structural high, partly subaerial landmass; light brown: platform, intermittently flooded; white: basin.

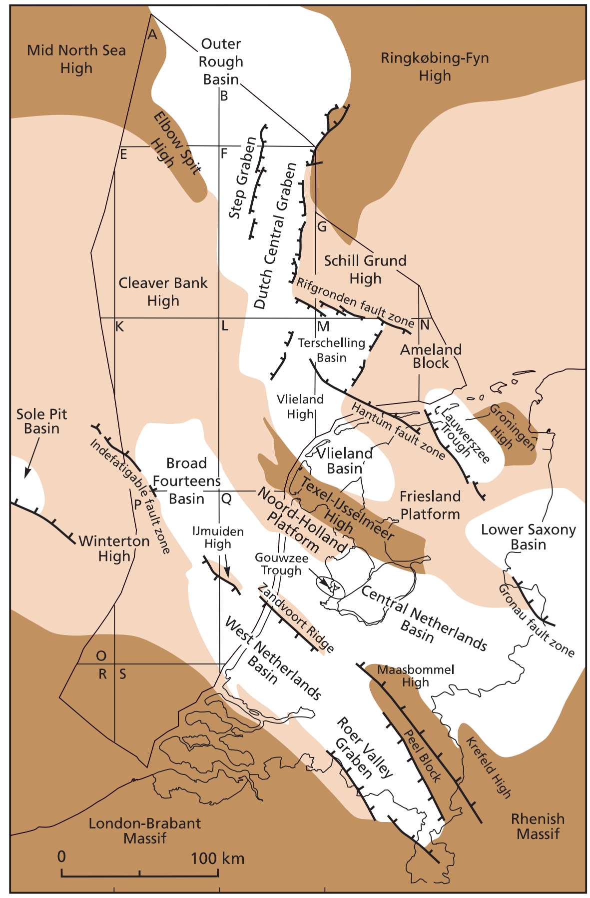


Fig. 3. Map of Late Jurassic to Early Cretaceous structural elements (after Van Adrichem Boogaert & Kouwe, 1993-1997). Shading as in Fig. 2.

an intricate pattern of smaller, fault-bounded basins and highs (Figs 2, 3; De Jager, this volume). According to Ziegler (1990), the change to this multi-basinal pattern is associated with the disintegration of Pangea. It occurred in the following extensional phases: Hardegsen (Scythian), Early Kimmerian (Anisian-Carnian), Mid Kimmerian (Aalenian-Callovian/Oxfordian) and Late Kimmerian (Kimmeridgian-Ryazanian). The intervals in between these phases were characterized by regional, thermal subsidence. The distinct partition between highs and basins was accentuated by halokinesis in the Permo-Triassic strata along existing structural trends while associated salt withdrawal often controlled the distribution of Jurassic depocentres.

During the late Middle and Late Jurassic, three major rift systems can be recognized (Fig. 3):

1. the N-S oriented Dutch Central Graben-Vlieland Basin system (including the Terschelling Basin),
2. the E-W oriented, Lower Saxony Basin system, extending into Germany,
3. a NW-SE block-faulted transtensional system comprising the Roer Valley Graben, West Netherlands Basin, Central Netherlands Basin and Broad Fourteens Basin, and also the Sole Pit Basin in British waters.

In the Early Jurassic, the area underwent a relatively constant subsidence and a uniform deposition of pelitic, open-marine sediments. Structural intricacy augmented gradually during the Early and Middle Jurassic and the pattern of structural elements shown in Figure 3 had been established by Callovian times during the Mid Kimmerian phase. Especially during the Late Jurassic each basin developed its own, specific, sedimentary history.

The Mid Kimmerian phase affected in particular the northern offshore area. Due to the thermal uplift of the Central North Sea Dome (Ziegler, 1990), the truncation of Middle Jurassic, Aalenian to Bathonian, strata was most rigorous in this area. The centre of the dome was located further north, mid-way between Scotland and Norway. From the Callovian to the Kimmeridgian, a mainly extensional tectonic regime associated with crustal thinning resulted in the rift system of the Dutch Central Graben. The uplift phase brought an end to the largely uniform subsidence and the sheet-like deposition of Lower and Middle Jurassic sediments. Rifting caused a differentiation into rapidly subsiding basins and more quiescent platform areas. This persisted into the Early Cretaceous. Outside the major basins, Upper Jurassic sediments are scarcely developed and usually associated with salt-induced rim synclines or transverse fault zones. Inside the basins, except perhaps locally in the West Netherlands Basin and Roer Valley Graben, the Upper Jurassic overlies, occasionally together with part of the Callovian, the older Jurassic strata unconformably. The hiatus involved represents as much as the time span Bajocian-Early Kimmeridgian, i.e more than 20 Ma, in the Dutch part of the Lower Saxony Basin.

During the Early Jurassic, the north-eastern area of the Netherlands was situated at the western edge of the German Ems Low (Fig. 2). This N-S oriented rift basin transformed in the Late Jurassic into the E-W trending Lower Saxony Basin (Ziegler, 1990). Toward the east, in Germany, the centre of the basin displays a complete Late Jurassic succession with hundreds of metres of evaporitic sediments of the Münder Formation (Fig. 1).

The Roer Valley Graben, West Netherlands Basin, Central Netherlands Basin and Broad Fourteens Basin were part of a different depositional province, together with the British Sole Pit Basin. Because Permo-Triassic evaporites are generally absent in this area, these basins are characterized by tilted half-grabens

Stratigraphy

The largely siliciclastic Jurassic sediments have been subdivided into the Altena, Schieland, Scruff and Niedersachsen groups (Fig. 4). The three latter groups represent mostly contemporaneous sediments that have been deposited in different basins. A volcanic formation, not assigned to a group, is known from one locality only. Since the Altena Group straddles the Jurassic-Triassic boundary into the Rhaetian, this Triassic stage will also be dealt with here. This is also the case for the Ryazanian, the lowermost Cretaceous stage, which includes sediments of the three other groups. A brief overview on a group, subgroup and formation level will be presented below. Many of the formations and members defined and described in detail by Van Adrichem Boogaert & Kouwe (1993-1997), will be dealt with under 'Sedimentary and structural development'.

Altena Group

The Altena Group consists mainly of argillaceous sediments with some calcareous intercalations in its lower part, and alternating calcareous and clastic deposits in its upper part. The group comprises the Sleen, Aalburg, Posidonia Shale, Werkendam and Brabant formations. Its present geographic distribution is confined to the Late Jurassic basins, where erosion and differential subsidence have resulted in thicknesses varying from a few metres to more than 1800 m (Fig. 5). The consistent, pelitic and marine lithofacies indicates that the absence of the group on the surrounding highs is mostly due to erosion. However, in places, considerable thinning towards the highs is apparent, e.g. against the London-Brabant Massif in the south. The most completely developed lithostratigraphic section is present in the north-western part of the Roer Valley Graben where it can attain a thickness of about 2000 m (Figs 5, 6). Scarce exposures of this group occur in the eastern Netherlands (Achterhoek). The depth of the base of this group is highly variable, reaching maxima of more than 5000 m in the Dutch Central Graben and the Broad Fourteens Basin (Fig. 7). The age of the group is Rhaetian to Oxfordian.

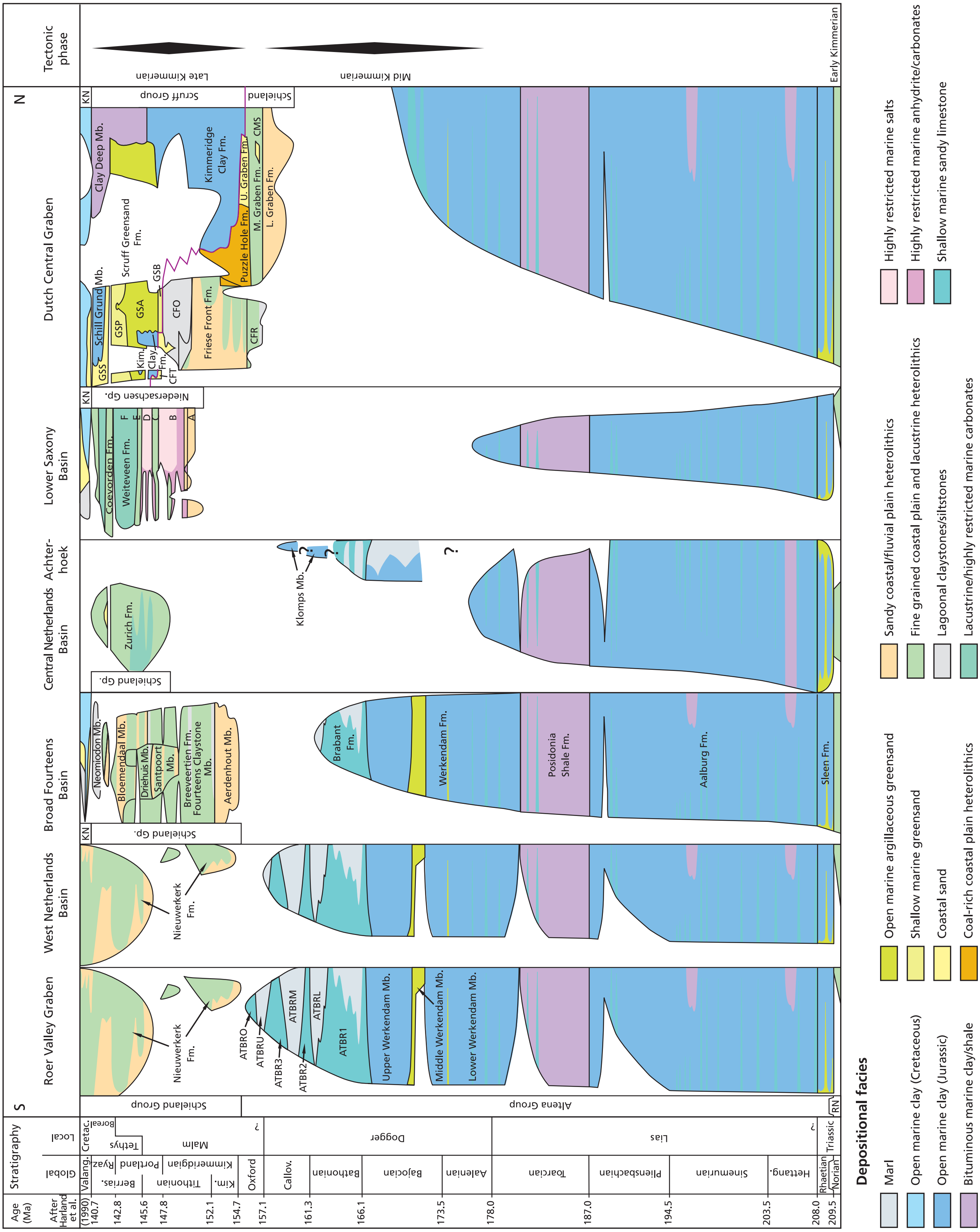
Schieland Group

The predominantly continental Schieland Group is subdivided into the Central Graben Subgroup and the Delfland Subgroup. This subdivision is mainly based on the basinal position of the strata. The Central Graben Subgroup comprises those formations of the Schieland Group that are confined to the Dutch Central Graben and Terschelling Basin. The Delfland Subgroup contains those that are present in the Vlieland Basin, Central Netherlands Basin, Broad Fourteens Basin, West Netherlands Basin and Roer Valley Graben. The age of the group ranges from Callovian to Barremian.

CENTRAL GRABEN SUBGROUP

The Central Graben Subgroup consists of alternations of sandstones, claystones and coal beds. Its deposition marks

Fig. 4. Stratigraphic framework of the Rhaetian to Ryazanian succession of the six larger Jurassic basins in the Netherlands (after Van Adrichem Boogaert & Kouwe, 1993-1997).
Abbreviated stratigraphic names:
Brabant Formation: ATBRO = Oisterwijk Limestone Mb; ATBRU = Upper Brabant Marl Mb; ATBR3 = Upper Brabant Limestone Mb; ATBRM = Middle Brabant Marl Mb; ATBR2 = Middle Brabant Limestone Mb; ATBRL = Lower Brabant Marl Mb; ATBR1 = Lower Brabant Limestone Mb.
Middle Graben Formation: CMS = Middle Graben Sandstone Mb.
Friese Front Formation: CFO = Oyster Ground Claystone Mb; CFT = Terschelling Sandstone Mb; CFR = Rifgronden Claystone Mb.
Weiteveen Formation: F = Serpulite Mb; E = Upper Marl Mb; D = Upper Evaporite Mb; C = Lower Marl Mb; B = Lower Evaporite Mb; A = Basal Clastic Mb.
Scruff Greensand Formation: GSS = Stortemelk Mb; GSP = Scruff Spiculite Mb; GSA = Scruff Argillaceous Mb; GSB = Scruff Basal Sandstone Mb.
KN = Rijnland Group.
RN = Upper Germanic Trias Group.



Age (Ma) After Harland et al. (1990)
Stratigraphy
Global
Local
S
N
Roer Valley Graben
West Netherlands Basin
Broad Fourteens Basin
Central Netherlands Basin
Achterhoek
Lower Saxony Basin
Dutch Central Graben
Tectonic phase
Late Kimmerian
Mid Kimmerian
Early Kimmerian
Schieland Group
Altena Group
Schieland Gp.
Scruff Group
Niedersachsen Gp.
Nieuwerkerk Fm.
ATBRO
ATBRU
ATBR3
ATBR2
ATBR1
ATBRM
ATBRL
Upper Werkendam Mb.
Middle Werkendam Mb.
Lower Werkendam Mb.
Neomiodon Mb.
Bloemendaal Mb.
Driehuis Mb.
Santpoort Mb.
Breeveertien Fm. Fourteens Claystone Mb.
Aerdenhout Mb.
Brabant Fm.
Werkendam Fm.
Posidonia Shale Fm.
Aalburg Fm.
Sleen Fm.
Zurich Fm.
Klomps Mb.
Coevorden Fm.
Weiteveen Fm.
Clay Deep Mb.
Schill Grund Mb.
Scruff Greensand Fm.
Kimmeridge Clay Fm.
Puzzle Hole Fm.
U. Graben Fm.
M. Graben Fm.
L. Graben Fm.
Friese Front Fm.
Kim. Clay Fm.
GSS
GSP
GSA
GSB
CFO
CFT
CFR
CMS
Depositional facies
Marl
Open marine clay (Cretaceous)
Open marine clay (Jurassic)
Bituminous marine clay/shale
Open marine argillaceous greensand
Shallow marine greensand
Coastal sand
Coal-rich coastal plain heterolithics
Sandy coastal/fluvial plain heterolithics
Fine grained coastal plain and lacustrine heterolithics
Lagoonal claystones/siltstones
Lacustrine/highly restricted marine carbonates
Highly restricted marine salts
Highly restricted marine anhydrite/carbonates
Shallow marine sandy limestone

the coastal-plain to shallow-marine phase in the subsiding Dutch Central Graben and Terschelling Basin after the Mid Kimmerian uplift and erosion. The subgroup is divided into five formations: Lower Graben, Middle Graben, Upper Graben, Puzzle Hole and Friese Front Formation. It interfingers with marine sediments of the Scruff Group. Due to both onlapping onto syndepositional highs and differential subsidence, the subgroup displays large variations in thickness. The thickest accumulation of approximately 2000 m has been recorded in the fault-bounded, north-eastern corner of quadrant F (Figs 8, 9). The age of the subgroup ranges from Callovian to Early Portlandian and its base is strongly diachronous (Herngreen & Wong, 1989).

DELFLAND SUBGROUP

The Delfland Subgroup consists of alternating sandstones and claystones with subordinate dolomites and coal beds. It contains three formations that were deposited in coastal-plain environments after the Mid Kimmerian uplift and erosion: the Zurich Formation in the Vlieland and Central Netherlands basins, the Breeveertien Formation in the Broad Fourteens Basin, and the Nieuwerkerk Formation in the West Netherlands Basin and Roer Valley Graben (Fig. 10). Depending on the structural setting, which differed from basin to basin, the thickness of the subgroup per basin is very variable. A maximum of 1500 m has been recorded in the Broad Fourteens Basin (Fig. 8). The age of the subgroup, also differing from basin to basin, ranges from Oxfordian to Barremian.

Scruff Group

This group consists of marine, locally bituminous claystones with thin intercalated carbonate beds, and of glauconitic, sometimes argillaceous, fine- to coarse-grained sandstones. The distribution of the group is limited to the Dutch Central Graben, Terschelling Basin and the northern part of the Vlieland Basin. The group is divided into the Kimmeridge Clay and Scruff Greensand formations. It interfingers with the largely continental Central Graben Subgroup. The great variation in thickness, especially apparent in quadrant F of the Dutch Central Graben, is due to strong differential subsidence and to erosion during the Subhercynian and Laramide inversion phases. A maximum of ca. 800 m is attained in the F3 quadrant (Figs 8, 9). The group's age ranges from Late Oxfordian to Ryazanian.

Niedersachsen Group

The Niedersachsen Group comprises predominantly claystones, but also includes marls, limestones and in its lower part evaporites. Bituminous shales occur locally. The lacustrine to hypersaline-lagoonal group is divided into the Kimmeridgian and Portlandian Weiteveen Formation and the Ryazanian Coevorden Formation (Fig. 11). Its distribution is restricted to the Lower Saxony Basin. In the Dutch part of this basin the group reaches a maximum thickness of just over 500 m (Fig. 8). Equivalent sediments in the German part of the basin include the Münder Formation and Bückeberg Formation (Fig. 1).

Zuidwal Volcanic Formation

This formation of massive volcanic rocks and brecciated volcanic agglomerates is known only from the Zuidwal volcanic centre under the Waddenzee (Van Bergen & Sissingh, this volume). It is overlain by the Vlieland Sandstone Formation of the Lower Cretaceous Rijnland Group. The Wadden Volcaniclastic Member (Zurich Fm, Delfland Subgp) may represent erosion products of the Zuidwal Volcano (Herngreen et al., 1991).

The relatively few occurrences of Jurassic igneous rocks elsewhere in the Netherlands are dealt with by Van Bergen & Sissingh (this volume).

Sedimentary and structural development

The Jurassic of the Netherlands comprises in essence two main depositional units. Both occur virtually only within the Late Jurassic basins. The lower unit corresponds to the Altena Group and can be considered as pre-rift and early syn-rift. Its base is locally unconformable, its top generally deeply eroded. This unit is of Rhaetian-Oxfordian age and represents the period from the end of the Early Kimmerian to the end of the Mid Kimmerian tectonic phase.

The upper unit represents the period between the latter phase and the Late Kimmerian (pulse II) phase (RGD, 1991). It is the main syn-rift sequence and consists of the Scruff and Niedersachsen groups and the bulk of the Schieland Group. Its age varies per basin within the range Middle Callovian to Ryazanian. It rests unconformably on the lower unit, except perhaps locally in the West Netherlands Basin and Roer Valley Graben, where the sedimentation seems to have been continuous. The corresponding hiatus varies from basin to basin and represents more than 20 Ma in the Dutch part of the Lower Saxony Basin. The unconformity involved is sometimes referred to as 'Mid Kimmerian Unconformity' (Burgers & Mulder, 1991; Van Adrichem Boogaert & Kouwe, 1993-1997). In some cases, poor stratigraphic control hampers its exact dating. Figure 4 gives an overview of the stratigraphic nomenclature and lithofacies for the main basins.

Pre-rift to early syn-rift unit: Altena Group (Rhaetian-Callovian/Oxfordian)

RHAETIAN-TOARCIAN

In many places the Altena Group overlies the Triassic Keuper Formation unconformably. Its Rhaetian to Toarcian part reflects the deposition between the Early and Mid

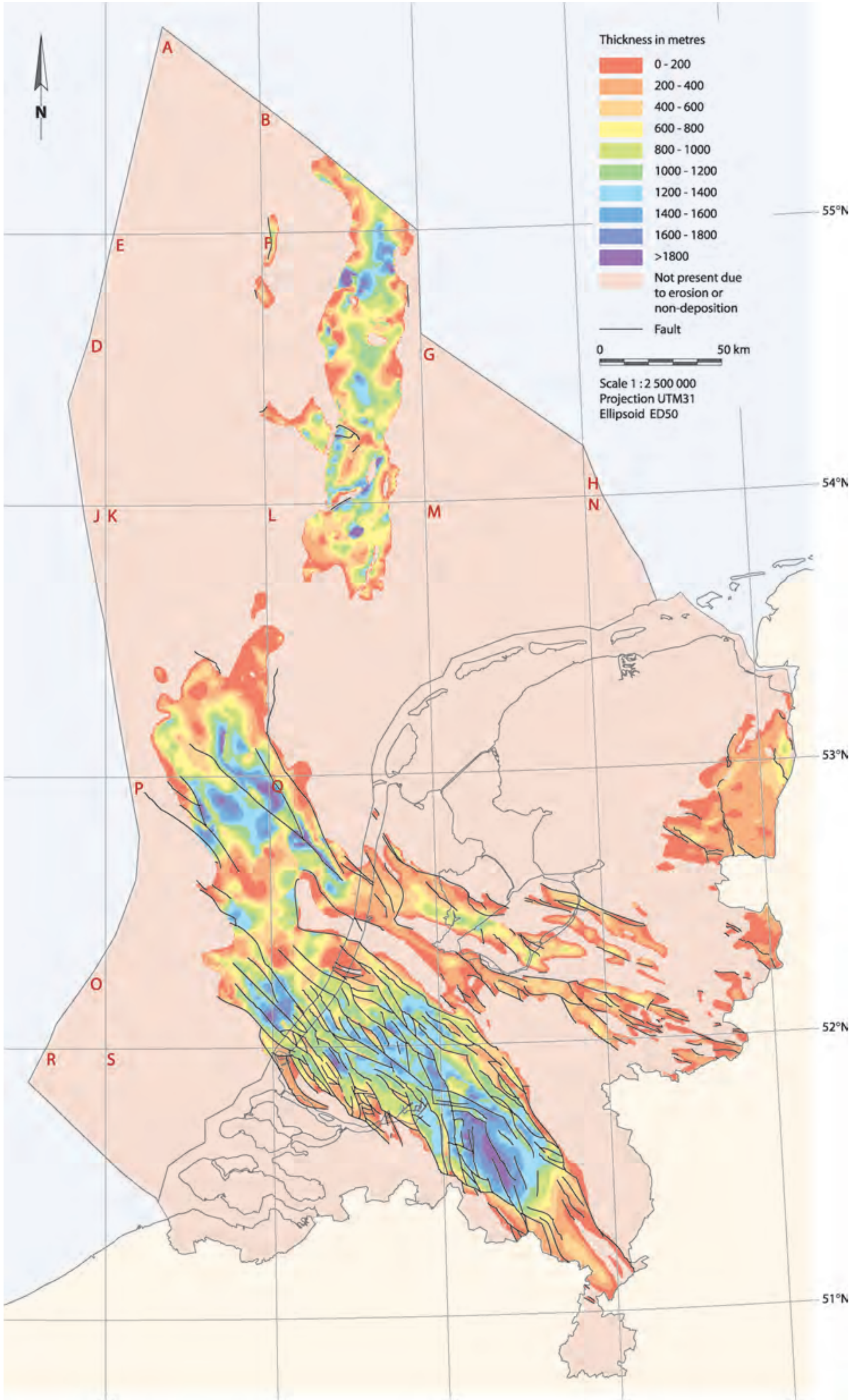


Fig. 5. Isopach map of the Altena Group (Duin et al., 2006).

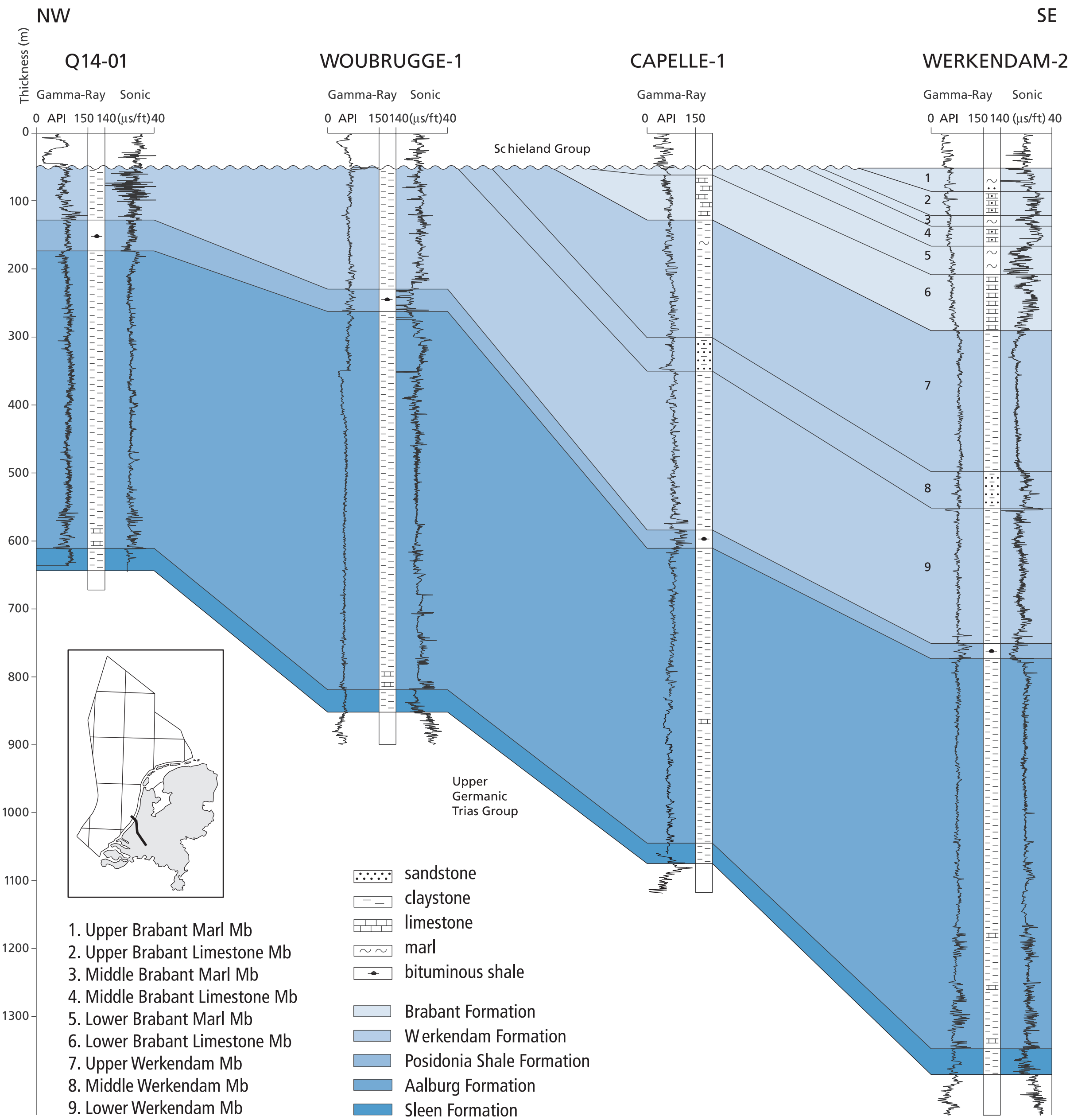


Fig. 6. Well-log correlation of the Altena Group in a NW-SE profile in the West Netherlands Basin (after NITG, 2002). Length of profile is ca. 70 km.

Kimmerian tectonic phases. This part comprises up to 1000 m of marine shales of a very uniform nature, which extend over large parts of north-west Europe.

Following the last pulse of the Early Kimmerian extensional phase in the earliest Rhaetian there was a marine transgression across large parts of Europe. The resulting sediments of the Dutch Sleen Formation, 20 to 45 m thick, are predominantly pelitic and contain lacustrine and marine fossils. This transgression lasted into the Early Jurassic, and during the Hettangian to Pliensbachian or earliest Toarcian the Aalburg Formation was deposited, consisting of a uniform section of up to 700 m of dark-grey to black claystones with abundant pyritized fossils. Towards the London-Brabant Massif, which formed the southern basin fringe during the Early Jurassic, the number of intercalated thin limestone beds increases.

Basin circulation became restricted during the Toarcian, causing anoxic bottom conditions to prevail over large parts of north-west Europe. A bituminous shale, generally ca. 30 m thick, the Posidonia Shale Formation, was deposited. Well-aerated conditions were re-established during the Late Toarcian when deposition of the Lower Werkendam Member started. This member consists of about

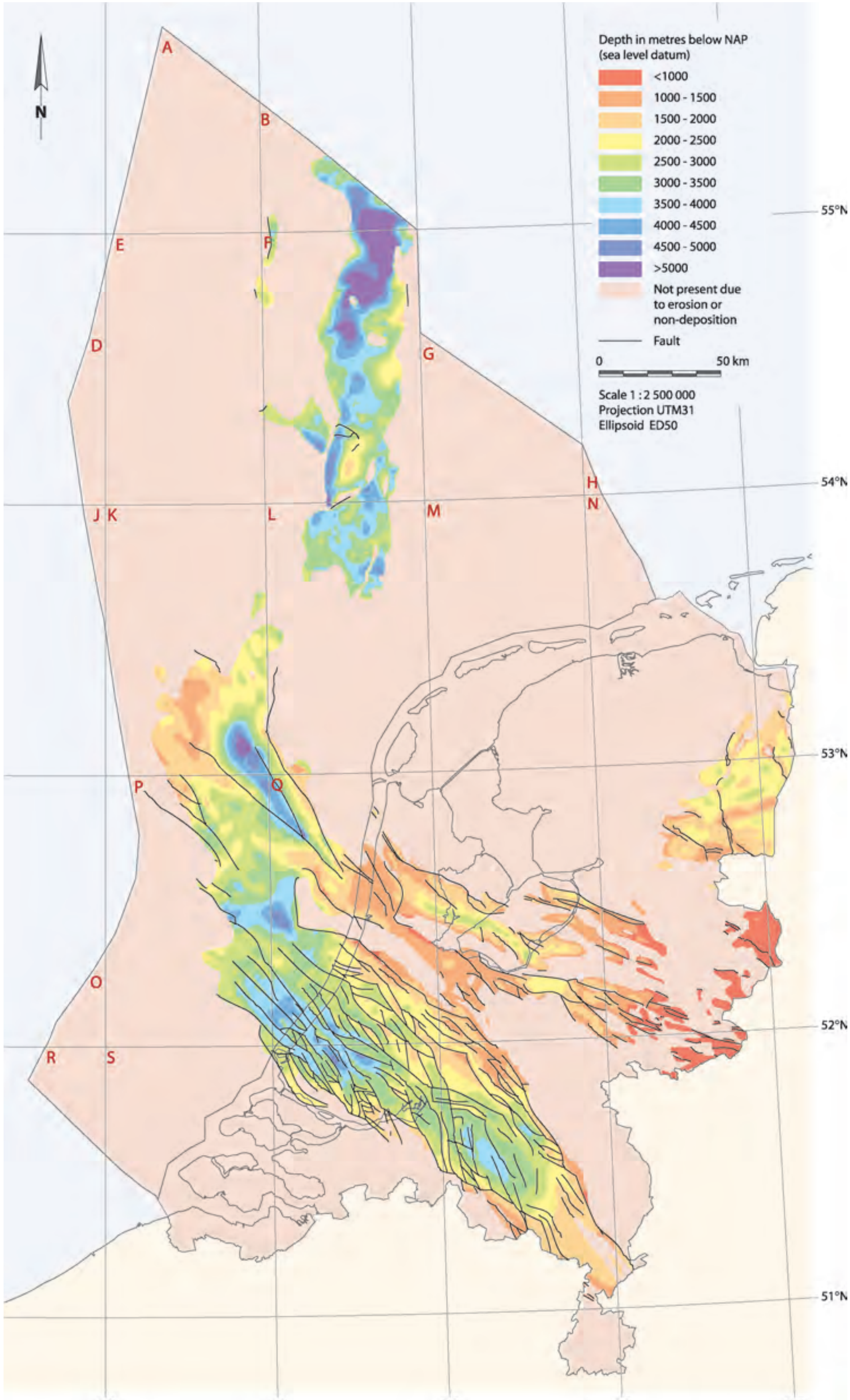


Fig. 7. Depth map of the base of the Altena Group (Duin et al., 2006).

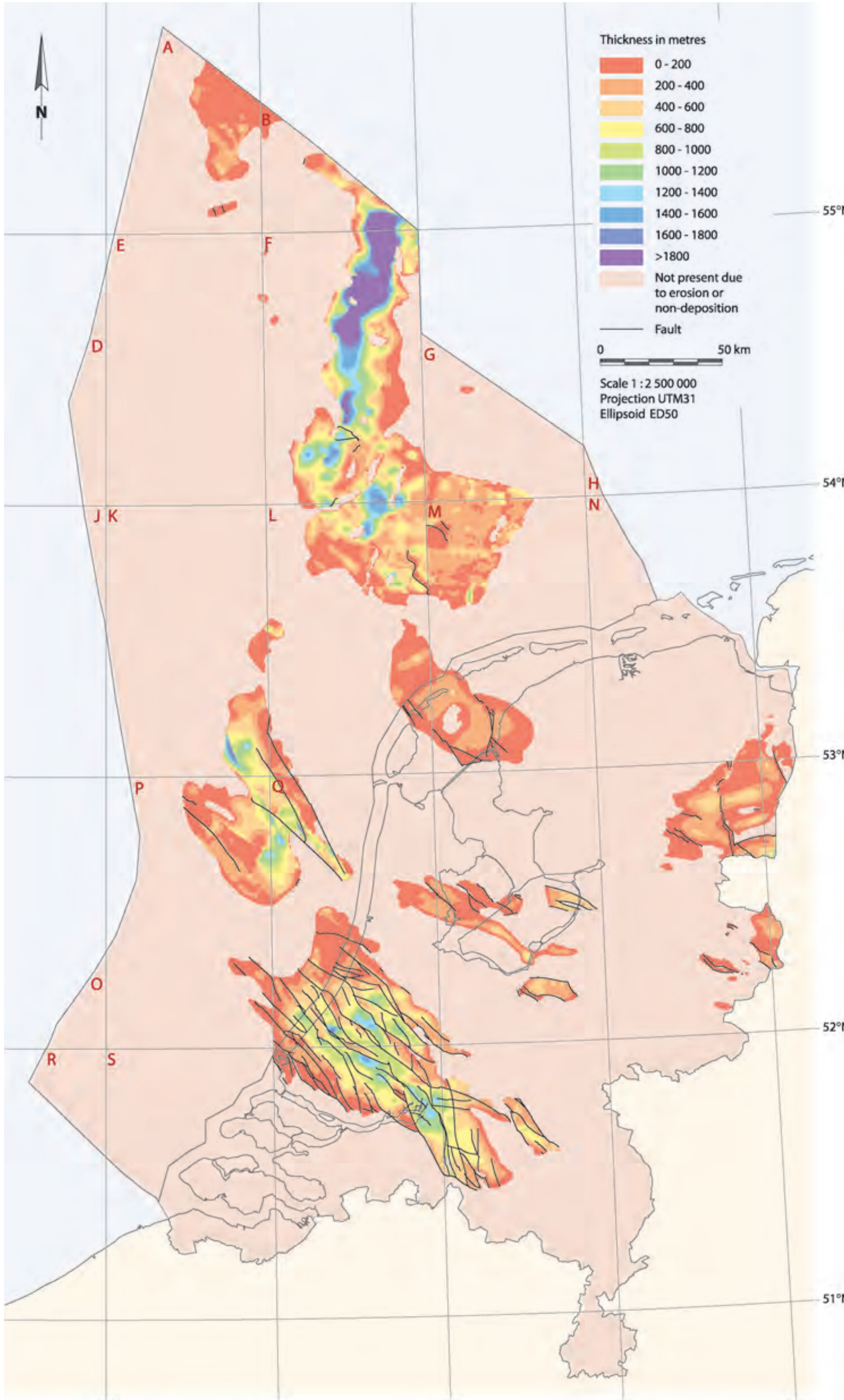


Fig. 8. Isopach map of the Schieland, Scruff and Niedersachsen groups in the various basins (Duin et al., 2006): Dutch Central Graben and Terschelling Basin (Central Graben Subgp, Delfland Subgp and Scruff Gp), Lower Saxony Basin (Niedersachsen Gp), Vlieland Basin (Central Graben Subgp, Delfland Subgp), Central Netherlands Basin, Broad Fourteens Basin, West Netherlands Basin and Roer Valley Graben (Delfland Subgp).

200 m of marine, silty mudstones with intercalated iron oolites and greensands.

Aalenian-Middle Callovian/Oxfordian

The upper part of the Altena group comprises up to 650 m of shallow-marine claystones, carbonates and sandstones. The sedimentation from the Aalenian (Fig. 12) to the Callovian and locally the Oxfordian was controlled by the Mid Kimmerian tectonic activity, which caused important structural differentiation. The principal effect was the thermal uplift of the Central North Sea Dome affecting the northern parts of the Netherlands and Germany (Pompeckj Block). The doming front gradually shifted southwards resulting in a considerable delay between the timing of rifting of the dome crest: mid-Aalenian in the central North Sea and Kimmeridgian in the southern Dutch provinces. The extent of intra-Jurassic truncation and non-deposition decreases away from this dome (Underhill & Partington, 1993). Consequently, the corresponding hiatus at the top of the Altena Group is most prominent in the northernmost basin, the Dutch Central Graben, where the open-marine sedimentation of the group terminated in the Bajocian. This area remained non-depositional from the Bathonian to Early Callovian.

While the erosion also affected the Terschelling Basin, Dutch Lower Saxony Basin and Central Netherlands Basin, this was not the case in the southern Broad Fourteens Basin, West Netherlands Basin and Roer Valley Graben. Deposition of the Werkendam Formation continued in the Bajocian, with a distinct influx of marine sands along the southern basin margins. The environment changed to shallow-marine, with the deposition of the sandy carbonates and marls or 'Cornbrash facies' of the Brabant Formation during the Bathonian. At least three carbonate–marl cycles were deposited in the Bathonian to Oxfordian. They are at present only preserved in erosional remnants in the southern basins, where a maximum thickness of 350 m can be attained.

In the Achterhoek, in the south-east of the Central Netherlands Basin, the depositional development until the Middle Bathonian shows significant similarities with the developments of the Roer Valley Graben and the German part of the Lower Saxony Basin (Herngreen et al., 1984). This entire region was part of one sedimentary basin which became differentiated only at the beginning of the Late Bathonian. From this time to the Callovian, an open-marine environment persisted in Germany and the Achterhoek, and claystones were deposited (Klomps mbr). On the contrary, deposition of shallow-marine sandy carbonates in the Roer Valley Graben and West Netherlands Basin persisted into the Oxfordian. In the Dutch part of the Lower Saxony Basin most of the upper part of the Altena Group appears to have been eroded.

Syn-rift unit: Schieland, Scruff and Niedersachsen groups (Middle Callovian/Oxfordian-Ryazanian)

Upper Jurassic to Lower Cretaceous deposits in north-west Europe show a step-by-step southward transgression. The transgressive periods alternated with partly tectonically controlled phases of short-term progradation of continental siliciclastics. This interaction is reflected by the repeated intertonguing of continental and marine sediments. The largely marine Scruff Group (Oxfordian-Ryazanian) and Rijnland Group (Early Cretaceous) interfinger with the mainly continental Schieland Group (Callovian-Barremian) and the lacustrine to hypersaline-lagoonal Niedersachsen Group (Kimmeridgian-Ryazanian; Fig. 4). Differential movement of fault blocks was caused by oblique slip in an overall transtensional regime and by halokinesis in areas with salts in the subsurface. Together with a high sediment input, this resulted in constantly shifting depocentres and in local thicknesses of the syn-rift unit of as much as 2500 m or more. Repetitive alternation of marine and continental deposits in this unit suggests that sedimentation was also influenced by eustatic sea-level movements. The successive, highly variable, depositional and tectonic patterns are discussed below; they are largely in line with the subdivision of Herngreen et al. (2003) into six 'stages'. It should be stressed that the time overlaps between the last four of these 'stages' reflect the diachronous character of relative sea-level fluctuations in the various basins.

Middle Callovian-Oxfordian

In the Middle and Late Callovian, predominantly continental sedimentation of the Schieland Group started along the axis of the northern Dutch Central Graben. The resulting basal, fluvial-plain deposits belong to the Lower Graben Formation. Due to onlapping onto syndepositional relief and to differential subsidence, this formation displays thickness variations from a few metres to about 560 m. This deposition gradually extended toward the southern part of the graben during the latest Callovian and the Oxfordian, and continued in the Kimmeridgian. First, the marginally marine Rifgronden Member of the Friese Front Formation was deposited. It indicates connections to open-marine areas in the northern Dutch Central Graben, or possible links with other marine realms further north.

The shift to the Middle Graben Formation (thickness up to 420 m) in the latest Callovian to Early Oxfordian reflects an instantaneous change to more fine-grained, lake and swamp-dominated sedimentation. This shift was accompanied by the short-lived, marine incursion of the Rifgronden Member in the southern Dutch Central Graben. This sea-level rise probably corresponds to a maximum-flooding surface as indicated by Haq et al. (1988). Dur-

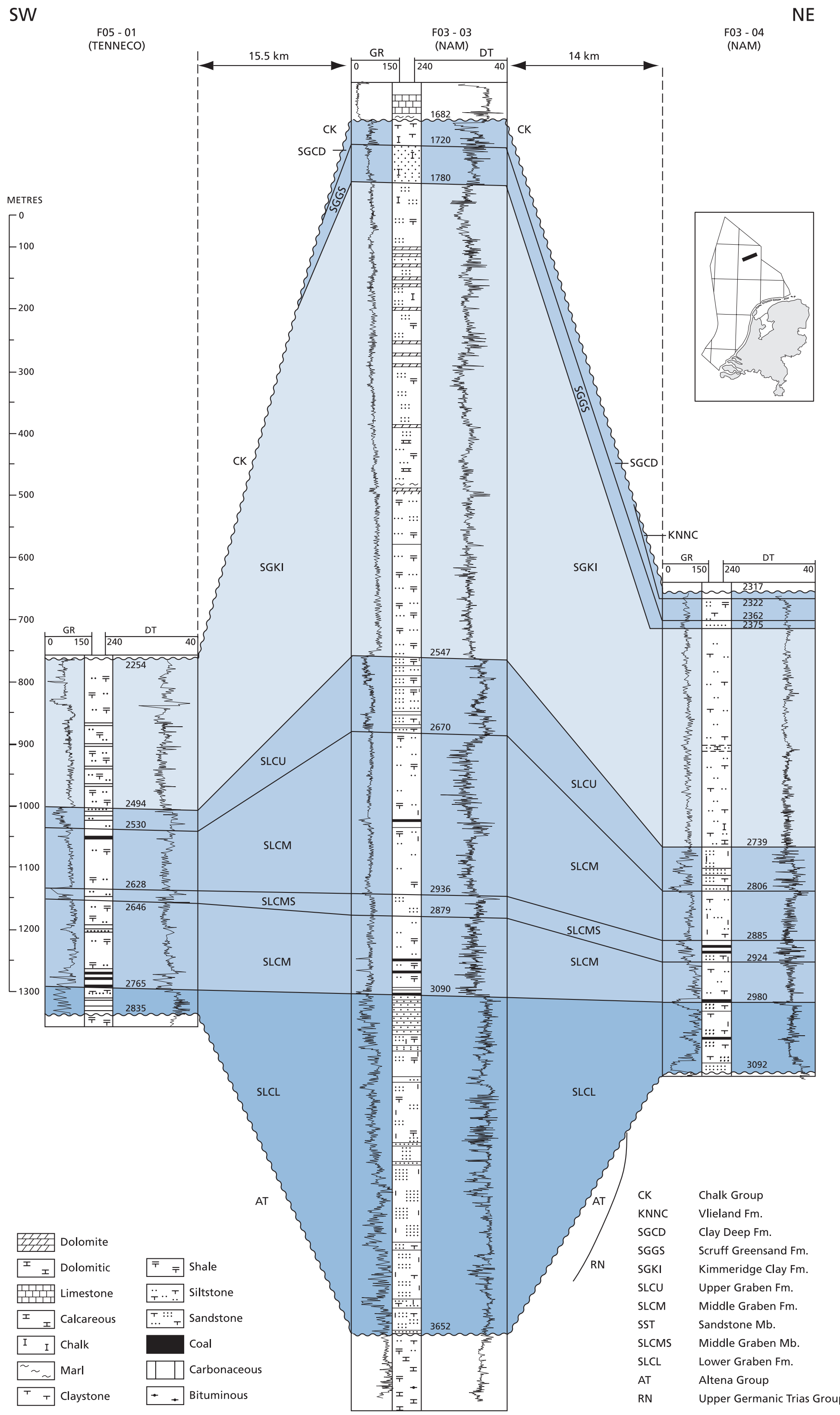


Fig. 9. Well-log correlation of the Central Graben Subgroup (SL codes) and Scruff Group (SG) in the Dutch Central Graben (after Herngreen & Wong, 1989, showing their stratigraphy).

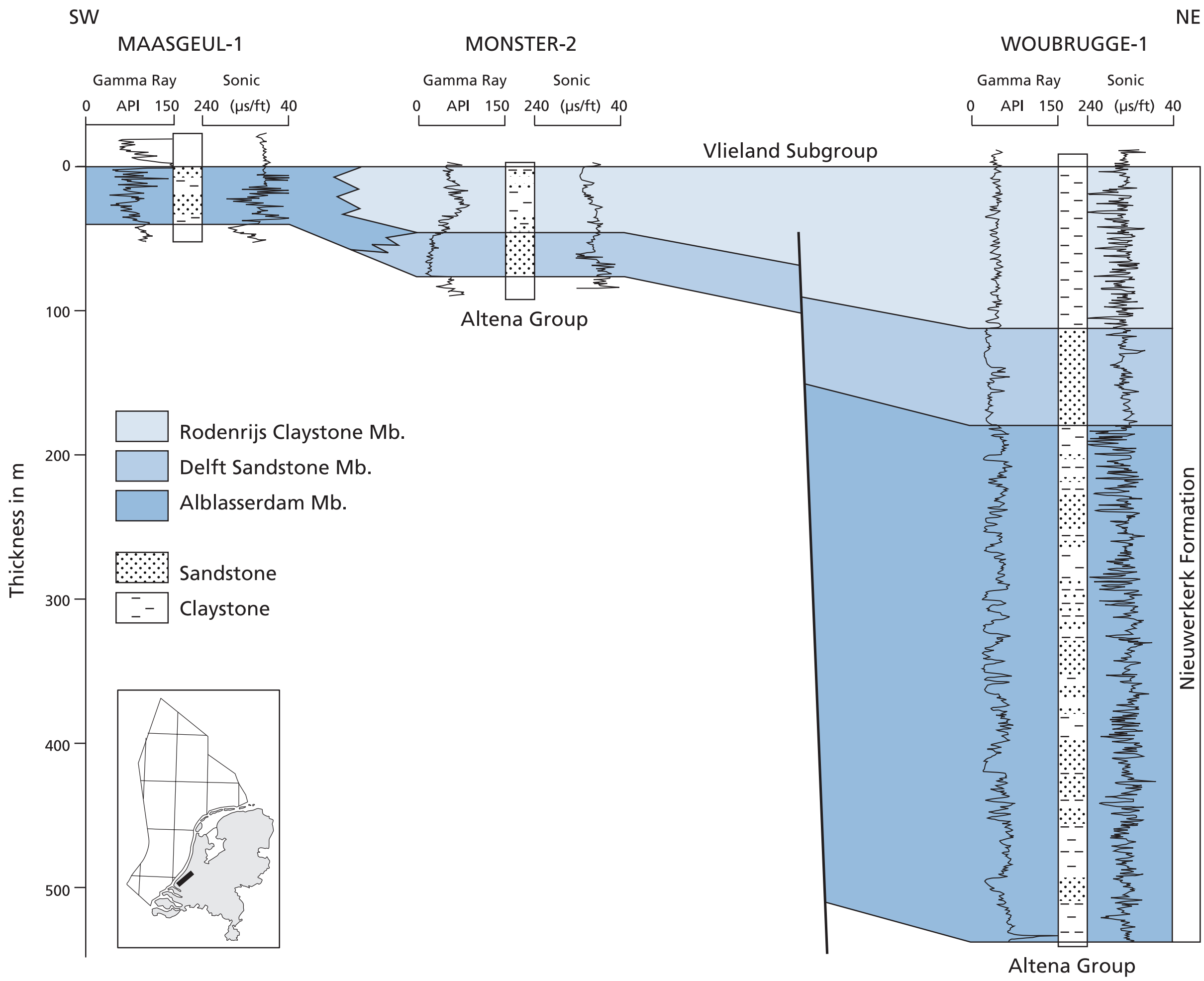


Fig. 10. Well-log correlation of the Nieuwerkerk Formation of the Delfland Subgroup in the West Netherlands Basin, showing that the thickness is strongly related to synsedimentary tectonics and subsequent erosion following inversion (after NITG, 2002). Length of section is ca. 65 km.

ing the Middle Oxfordian, lacustrine conditions prevailed throughout the Dutch Central Graben. In the Late Oxfordian, occasional marine incursions reached the extreme north of the area.

The sea penetrated southwards into the Dutch Central Graben, first into block F3 during the Middle and Late Oxfordian. Originally, a stacked, prograding coastal-barrier sand complex, the Upper Graben Formation (thickness up to 125 m), developed, while south of it a paralic delta-plain formed, represented by the Puzzle Hole Formation in the eastern central part of the F quadrant. Especially in rapidly subsiding rim synclines around salt diapirs, this formation can reach a thickness of about 400 m (Fig. 13). North of the barrier complex, the contemporaneous open-marine facies is represented by the Kimmeridge Clay Formation. The complex became submerged at the end of the Oxfordian. At this instance, the sea also transgressed into the Step Graben, depositing Kimmeridge Clay on the Upper Permian Zechstein Group.

Non-marine Schieland Group deposition had already started in the Dutch Central Graben during the Callovian, whilst deposition of the marine Brabant Formation continued into the Early and Middle Oxfordian in the southern Netherlands (Haanstra, 1963; NAM & RGD, 1980). The Late Kimmerian I pulse during the earliest Kimmeridgian ended the marine Altena Group deposition in the Roer Valley Graben, West Netherlands Basin and Broad Fourteens Basin, and probably also in the Central Netherlands Basin and Lower Saxony Basin. In the Roer Valley Graben and West Netherlands Basin, the Brabant Formation is locally overlain unconformably by erosional remnants of terrestrial, Late Oxfordian to Portlandian deposits of the Schieland Group (Nieuwerkerk Fm).

Early Kimmeridgian

During the Kimmeridgian the northern part of the Dutch Central Graben evolved into a principal depocentre which received a thick section of increasingly marine sediments. By contrast, the deposition in the Broad Fourteens Basin,

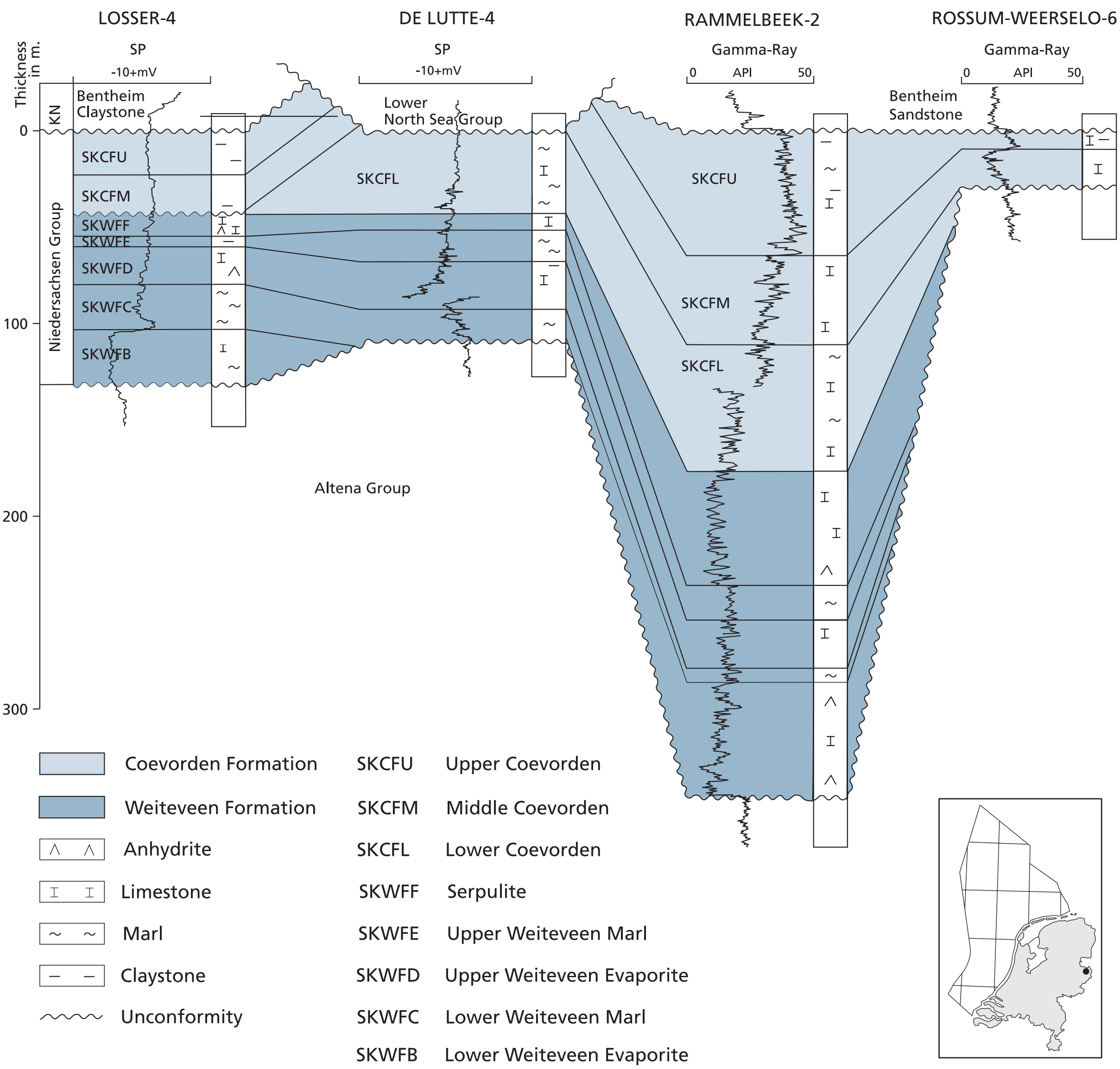


Fig. 11. Well-log correlation of the Niedersachsen Group in the Dutch part of the Lower Saxony Basin (after NITG, 1998). KN: Rijnland Group.

the Dutch part of the Lower Saxony Basin and the Vlieland Basin resumed in an environment of floodplains and lakes, with occasional sand-supplying rivers and marine incursions (Fig. 14). In the Vlieland Basin a volcanic event took place between the Oxfordian and the Portlandian (143 to 156 Ma) as is witnessed by the Zuidwal volcanic dome and associated volcanic rocks (Perrot & Van der Poel, 1987; Herngreen et al., 1991; Van Bergen & Sissingh, this volume).

Under the continuing transgression, the paralic conditions of the Puzzle Hole Formation in the Dutch Central Graben changed to the open-marine environment of the Kimmeridge Clay Formation, which reaches in places a thickness of 760 m. The thin coastal-barrier sands that occur locally in the uppermost parts of the Puzzle Hole Formation suggest the presence of a backstepping coastal-barrier system. Subhercynian and Laramide erosion, however, has obscured the character of this transgression. In the southern part of the F quadrant, the predominantly paralic environment of the Puzzle Hole Formation had given way to the coastal plain of the Friese Front Formation at the end of the Oxfordian. In the Early Kimmeridgian a transgressive pulse once more expanded the depositional area in the southern part of the Dutch Central Graben. Seismic and well data suggest that the boundaries between the depositional areas of the Puzzle Hole, Kimmeridge Clay and Friese Front formations were fault-controlled.

In the Broad Fourteens Basin, widespread sedimentation started in the Early Kimmeridgian with the sandy fluvial-plain deposits of the Aerdenhout Member of the

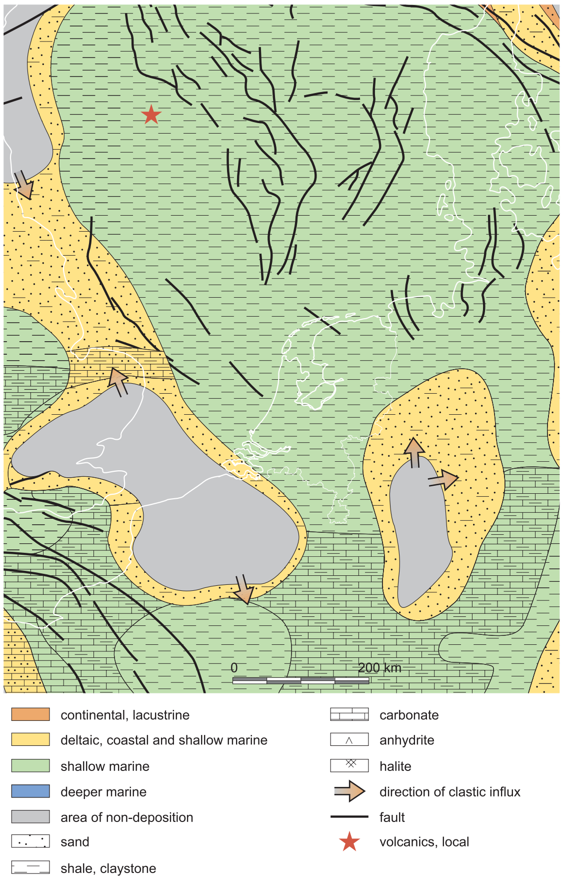


Fig. 12. Paleogeographic map of the Netherlands and adjacent areas during the Sinemurian-Aalenian (after Ziegler, 1990). Legend also applies to Fig. 14.

Breeveertien Formation. This member unconformably overlies the Altena or Upper Germanic Trias groups. The progressing transgression as noted in the Dutch Central Graben, coincides in the Broad Fourteens Basin with a shift from fluvial-plain (Aerdenhout Mbr) to lacustrine and lagoonal deposition (Fourteens Claystone Mbr). Deposition of the basal Weiteveen Clastic Member in the Dutch part of the Lower Saxony Basin is considered to have started at the end of the Early Kimmeridgian.

The appearance of coarser sediments in the Dutch Central Graben (main part of Friese Front Fm, Puzzle Hole Fm and Upper Graben Fm), the Broad Fourteens Basin (Breeveertien Fm, Aerdenhout Mbr) and the Dutch part of the Lower Saxony Basin (Weiteveen Basal Clastics Mbr) corresponds with the diachronous beginning of the Late Kimmerian phase (Haanstra, 1963; 't Hart, 1969).

The ensuing uplift ended the first depositional phase of the Nieuwerkerk Formation in the rapidly subsiding Roer Valley Graben and West Netherlands Basin. This syn-rift deposition was dominated by braided river channels, with rapid lateral variations in sand-body thickness. The main transport direction was to the north-west, parallel to the newly formed graben boundary faults. The main sediment-source area may have been the Ardennes and Rhenish Massif (Den Hartog Jager, 1996).

Late Kimmeridgian-earliest Portlandian

In the Late Kimmeridgian, the transgression led to the first marine sedimentation in the southern Dutch Central Graben and the Terschelling Basin, as indicated by the presence of the Oyster Ground Claystone Member (Friese Front Fm). Two transgressive phases took place. The first was already in the early Late Kimmeridgian when it reached block F18, while L2 was inundated during the latest Kimmeridgian. The second flooding was associated with onlap onto Triassic and Permian strata in the Terschelling Basin (Oyster Ground Claystone Mbr in F15) and the Vlieland and Central Netherlands Basins (continental Zurich Fm). Simultaneously, the depocentre of the Dutch Central Graben shifted from the eastern graben axis (between F3 and L3) to the western graben margin (between F2 and L2). This was caused by tectonic tilting and associated halokinesis.

In the West Netherlands Basin and Roer Valley Graben, lacustrine and fluvial-plain deposition was confined to those fault blocks that underwent strongest subsidence. This deposition terminated due to uplift in the Kimmeridgian. The non-deposition lasted until the latest Portlandian-Ryazanian.

Portlandian

The Early Portlandian transgression resulted in deposition of the Terschelling Member of the Friese Front Formation in coastal settings along the southern fringe of the Dutch Central Graben (in L9 and L12). To the north this formation grades into the open-marine Scruff Greensand Formation. Thick (up to 360 m), sand-dominated sections were deposited (F15 and F18), probably in topographic depressions on downfaulted graben margins and/or in salt-induced rim synclines (Herngreen & Wong, 1989). Equivalent, thinner sandstone beds are found in the eastern Terschelling Basin, the Vlieland Basin and in blocks F3 and F5 of the Dutch Central Graben. During the Portlandian the Vlieland Basin became divided into two sub-basins by the Zuidwal volcanic dome (Herngreen et al., 1991). During the Early Portlandian, marine conditions prevailed in the northern subbasin (Kimmeridge Clay Fm) while lagoonal to lacustrine conditions dominated in the southern subbasin (Zurich Fm).

The Scruff Greensand Formation grades northward into the Kimmeridge Clay Formation in the northern Dutch Central Graben. The depocentre of the latter formation, which was previously situated in the northern part of the area, now shifted to the southernmost part of the graben and the northern Vlieland Basin. In the southern B quadrant (and in the UK), basin circulation stagnated, resulting in deposition of the bituminous claystones known in the

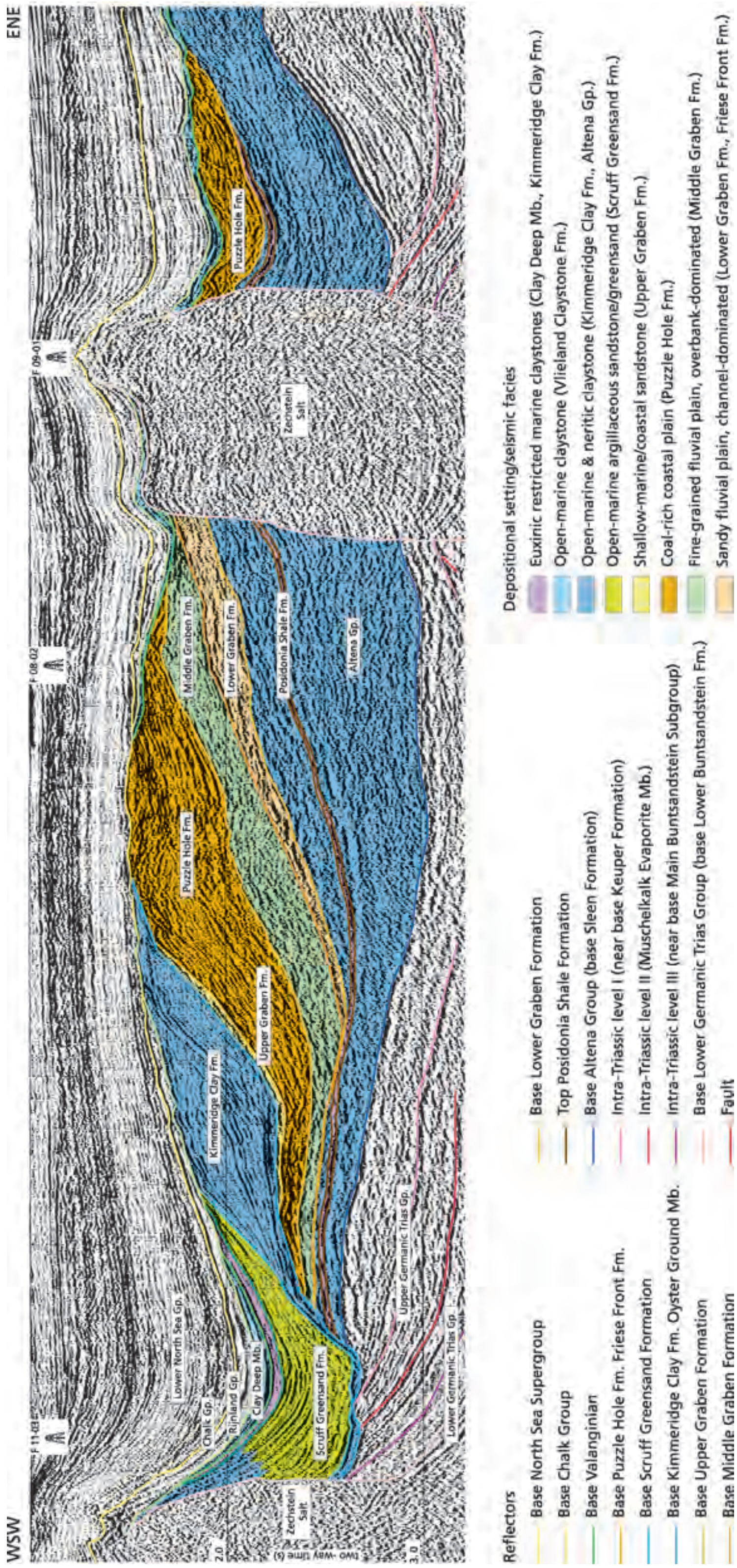


Fig. 13. Seismic section across the Dutch Central Graben. The Jurassic-Lower Cretaceous section has been colour-coded according to the main depositional facies. Halokinetic movements have strongly influenced Jurassic sedimentation, as indicated by thickening of the Altena Group near well F08-02 and of the Scruff Greensand Formation near F11-03 (after Van Adrichem Boogaert & Kouwe, 1993-1997). Length of profile is ca. 75 km.

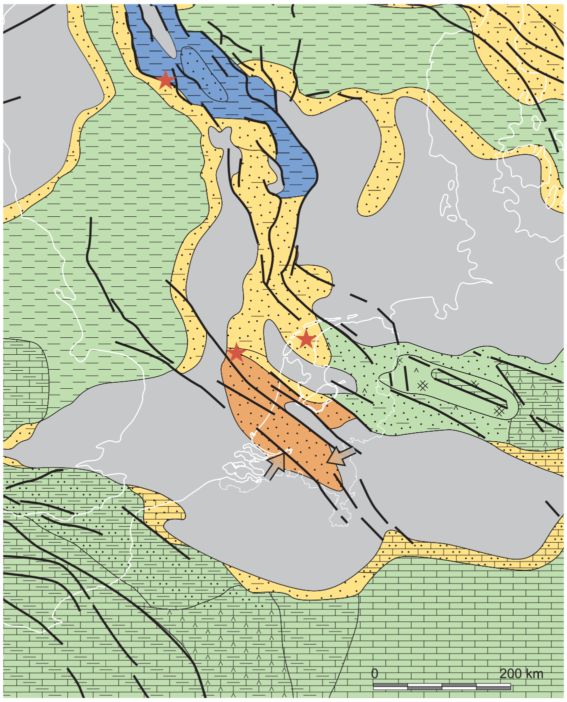


Fig. 14. Paleogeographic map of the Netherlands and adjacent areas during the Kimmeridgian-Tithonian (after Ziegler, 1990). For legend see Fig. 12.

Netherlands as the Clay Deep Member. This member represents one of the southernmost occurrences of the Kimmeridge 'hot shale' facies of Barnard & Cooper (1981) in the North Sea area. In the Dutch Central Graben the Clay Deep Member looses its bituminous character in southern direction (Herngreen & Wong, 1989).

During the Late Kimmeridgian and Early Portlandian the deposition of clastic sediments of the Weiteveen Formation in the Lower Saxony Basin was intermittently replaced by accumulation of evaporites and carbonates (Fig. 11). Possibly these cycles may be correlated with clastic and occasionally evaporitic cycles of the Breeveertien Formation in the Broad Fourteens Basin. The lithofacies (carbonates, marls, evaporites and coals) of the Zurich Formation in the Central Netherlands Basin are identical to those in the Niedersachsen Group in the Dutch part of the Lower Saxony Basin.

Late Portlandian-Ryazanian

In the Late Portlandian the sea briefly reached the Vlieland Basin depositing the Scruff Greensand Formation. In the southern Dutch Central Graben it became shallower, inducing the northward progradation of shallow-marine, glauconitic greensands with abundant sponge spicules (Scruff Spiculite Mbr). Subsequently, the sea occasionally inundated also the more elevated areas (e.g. salt domes) of the Central Netherlands Basin and the Dutch part of the Lower Saxony Basin. In the first basin the depositional area expanded to the north-west.

In the Broad Fourteens Basin, there is a transition from lacustrine and lagoonal coastal-plain muds of the Fourteens Claystone Member and the Driehuis Mottled Claystone Member to predominantly fluvial-plain deposits of the Bloemendaal Member. In the Lower Saxony Basin, this shoaling tendency is demonstrated by the Serpulite Member of the Weiteveen Formation, a widespread carbonate-rich deposit that marks the peak of clastic starvation in this region. The 'Serpulite' facies was replaced by argillaceous lacustrine deposition of the Coevorden Formation at the beginning of the Ryazanian. This coincides with the transition from the sandy fluvial Bloemendaal Member to the lagoonal Neomiodon Claystone Member in the Broad Fourteens Basin. This episode is marked by the widespread deposition of the Nieuwerkerk Formation in the Roer Valley Graben and the West Netherlands Basin (Van Adrichem Boogaert & Kouwe, 1993-1997; DeVault & Jeremiah, 2002). Up to 1500 m of, locally coarse, fluvial-plain sediments were deposited in rapidly subsiding grabens and half-grabens of these basins (Fig. 10). The basin margins were either covered by thin, condensed successions, or remained exposed and non-depositional.

Late Ryazanian

During the Ryazanian the stagnant marine basin area in the northern Dutch Central Graben extended southward into block F5. Deposition of the Clay Deep Member completely superseded that of the Scruff Greensand Formation around mid-Ryazanian time. This change in deposition coincided with the first activities of the Late Kimmerian II pulse. This pulse also caused local truncation of the Scruff Greensand and Friese Front formations in the southern part of the graben. Initially, sedimentation of the Scruff Greensand Formation resumed briefly, giving rise to the Stortemelk Member. Subsequently, the marine basin expanded markedly southwards, and Kimmeridge Clay Formation (Schill Grund Mbr) was deposited in this previously shallow-marine to continental area. This marine incursion also reached the Vlieland Basin, where the Stortemelk Member is intercalated with lagoonal to lacustrine deposits of the Zurich Formation. In the Broad Fourteens Basin the Late Kimmerian II pulse is expressed as two minor unconformities (base and intra-Neomiodon Claystone Mbr, Breeveertien Fm).

In the Dutch Central Graben, large-scale differential subsidence ended with this pulse (Late Ryazanian). From then on, a more uniform sedimentation started, associated with the post-rift thermal sag phase. This is demonstrated by the contrast between the highly variable thickness of the sequence underlying the Stortemelk Member and the more uniform development of the post-uplift succession (Clay Deep, Stortemelk and Schill Grund members).

Economic geology

Various Jurassic and Ryazanian sediments have economic significance in the Netherlands, either as reservoir rock or as source rock for hydrocarbons (De Jager & Geluk, this volume). The Posidonia Shale Formation is the main oil-source rock in the Dutch subsurface (Bodenhausen & Ott, 1981). Both marine algal sapropel and land-derived organic matter are present (type II kerogen). The relatively small oil occurrences in most of the Netherlands may be attributed to the patchy distribution of the formation, which resulted from severe erosion during later tectonic phases. Additional oil-source rocks exist in the Coevorden Formation in the Lower Saxony Basin. They generated the oil of the relatively large Schoonebeek field (De Jager & Geluk, this volume). This horizon is also a known oil-source rock in the German part of the basin, where it belongs to the Bückeberg Formation (Fig. 1).

Moreover, the Lower Jurassic Aalburg Formation with its type II source-rock characteristics, probably contributed to some degree to oil accumulations in the Dutch Central Graben (Wong et al., 1989) and West Netherlands Basin (De Jager et al., 1996). The bituminous Clay Deep Member of the Kimmeridge Clay Formation in the northern part of the Dutch Central Graben also has oil-generating potential. However, the areal and stratigraphic extent of this member are much smaller than those of equivalent, very productive, hot shales in the Kimmeridge Clay in the British, Danish and Norwegian sectors of the North Sea.

Coal-bearing strata in the Puzzle Hole, Middle Graben and Friese Front formations may locally have generated gas. However, the burial of these coals is considered insufficient to have yielded economic quantities (Wong et al., 1989).

Accumulations of oil and locally gas have been discovered in the sandstones of the Lower Graben Formation, Upper Graben Formation and Friese Front Formation in the Dutch Central Graben area. In the Broad Fourteens Basin, oil and gas have been found in sandstone intervals of the Breeveertien Formation. In the West Netherlands Basin, various members of the Brabant and Nieuwerkerk formations are locally oil and gas-bearing.

The traps in the Dutch Central Graben area are related to salt tectonics, and are mainly present as four-way dip closures in turtle-back anticlines. Some of the reservoir rocks in this area are strongly fractured. Elsewhere, combinations of fault and dip-closed structures are present. The top seals are various Jurassic shaly intervals.

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Cretaceous

G.F.W. Herngreen & Th.E. Wong

ABSTRACT

The Cretaceous in the Netherlands is represented mainly by the marine, Lower Cretaceous Rijnland Group (up to 1400 m thick in places) and Upper Cretaceous to lowermost Paleocene Chalk Group (locally more than 1800 m thick). During the Early Cretaceous, the rifting tectonics of the Jurassic changed into regional subsidence. As a result of rising sea level, the sedimentation, previously restricted to the basins, extended over the adjacent platforms and highs. This long-term transgression was not continuous: subordinate sea-level fluctuations resulted in various transgressive basal sands or prograding coastal-barrier sands of the Vlieland Sandstone Formation of the Rijnland Group. Meanwhile, continental siliciclastics of the Schieland Group (Late Jurassic to Barremian) continued to be deposited locally at the southern, landward side of the Rijnland Group. By Aptian times, deeper marine conditions resulted in the mainly marly deposits of the Holland Formation of the same group. Various unconformities occur within this group. A considerable hiatus reflects the Austrian tectonic phase and subsequent Albian transgression. During the Late Cretaceous, the influx of fine-grained clastics into the marine realm diminished. A fairly uniform succession of marls and limestones of the Texel and Ommelanden formations of the Chalk Group developed. This period of calm sedimentation was interrupted by the Subhercynian and Laramide inversion phases, of Santonian-Campanian and Paleocene age respectively. The resulting erosion removed much of the Cretaceous in the former, Jurassic basinal areas. The Cretaceous/Tertiary boundary is exposed in southern Limburg. Numerous oil and gas fields produce from Lower Cretaceous sandstones. In contrast, only one oil field and one gas field produce from the Chalk Group. A few quarries and aquifers in Cretaceous rocks locally yield limestone for cement production, and water, respectively.

Keywords: Danian, Netherlands, stratigraphy, basin development, tectonic inversion

Introduction

Previous research

The Cretaceous geology of the Netherlands was summarized by Van Staalduinen et al. (1979). Since then, numerous publications devoted to the various Cretaceous basins have appeared; in these, tectonic aspects were more and more incorporated. Special mention deserve i) Burgers & Mulder (1991) and Dronkers & Mrozek (1991), who dealt with the structural development, ii) the revised lithostratigraphic nomenclature compiled by Van Adrichem Boogaert & Kouwe (1993-1997), and iii) the 'Geological Atlas of the Subsurface of the Netherlands', an onshore mapping project (scale 1 : 250 000) recently completed with the publication of a compilation at scale 1 : 1000 000 by the Netherlands Institute of Applied Geoscience TNO (NITG, 2004). These publications resulted mainly from the exploration for oil and gas. Based on TNO data, Duin et al. (2006) recently published a series of thickness and depth maps, some of which have been used in this chapter.

Geological setting

The remarkable differences in sedimentary development of the various basins around the Jurassic/Cretaceous boundary can only be understood against the background of tectonic events in the Northwest European Basin, more particularly Kimmerian rifting in the North Sea area (Fig. 1; De Jager, this volume). Early in Middle Jurassic times the thermal Central North Sea Dome came into existence, which caused extensive Mid-Kimmerian uplift and subsequent deep erosion of older strata. In Callovian to Ryazanian times, increased heat flow, magmatic activity with anorogenic volcanism and extensional faulting characterized Late Kimmerian tectonic pulses. Rift structures like the Dutch Central Graben, Broad Fourteens Basin, West Netherlands Basin – Roer Valley Graben, Central Netherlands Basin, Lower Saxony Basin, Vlieland and Terschelling basins, and Lauwerszee Trough (Fig. 2) originated during this period or constitute rejuvenated Paleozoic elements. In the Early Cretaceous, thermal subsidence became important, and starting in the Valanginian, transgression took gradually place. This flooding was briefly interrupted by the Early Albian, Austrian tectonic phase and followed by the 'Albian transgression' which is widely recognized in northwest Europe (Crittenden, 1987). In Late Cretaceous times, the influx of clastics gradually diminished due to the submergence of nearby source areas, and calcareous sedimentation increased. The Subhercynian (Santonian-Campanian) and Laramide (Paleocene) tectonic phases caused strong differential subsidence and tectonic inversion of former Jurassic basins. During these phases, older, in part Jurassic, normal faults were reactivated as reverse faults and much of the inverted basin fills were eroded (Fig. 3). Both phases are proba-

Chronostratigraphy		Group	Tectonic phase	Events
Early Tertiary	Paleocene	Lower North Sea	Laramide	Renewal of inversion tectonics; local erosion, minor rifting. Widespread carbonate sedimentation ends at end Danian.
Late Cretaceous	Maastrichtian Campanian Santonian Coniacian Turonian Cenomanian	Chalk	Subhercynian	Strong inversion and erosion of Upper Jurassic depocentres; regional differential subsidence. Onset of widespread carbonate sedimentation.
Early Cretaceous	Albian Aptian Barremian Hauterivian Valanginian Ryazanian	Rijnland Schieland/ Niedersachsen (Scruff)	Austrian Late Kimmerian Phase II	Unconformity; local uplift, followed by erosion and regional Albian transgression. Onset of Cretaceous transgression over 'Late Kimmerian Unconformity'. Diminishing rift activity; thermal subsidence.
Late Jurassic			Late Kimmerian Phase I	Major rifting and differential subsidence; rapid sedimentation over 'Mid Kimmerian Unconformity' in narrow and restricted basins; local volcanism.

Fig. 1. Tectonic phases and main events from Late Jurassic to earliest Tertiary in the Netherlands (after Ziegler, 1990).

bly related to Alpine foreland compression (Ziegler, 1982, 1990); however, Baldschuhn et al. (1991) considered the inversion to have been the result of strong, local compression, and Kockel (2003) concluded that basic questions on the driving mechanism of inversion remain unsolved. By the end of the Cretaceous also the London-Brabant and Rhenish massifs in the south became submerged.

Stratigraphy

This chapter follows the standard subdivision of the Cretaceous (Gradstein et al., 2004) with the exception that the name Ryazanian is applied for the lowest Cretaceous stage (cf. Herngreen & Wong, 1989; Hoedemaker & Herngreen, 2003). The Ryazanian strata in the Netherlands are dealt with in the chapter 'Jurassic' (Wong, this volume). The Cretaceous of the Netherlands above the Ryazanian consists mainly of two lithostratigraphic groups (Van Adrichem Boogaert & Kouwe, 1993-1997). Both groups are marine. They are the Lower Cretaceous, siliciclastic Rijnland Group and the Upper Cretaceous, and also Danian, largely carbonate Chalk Group (the Paleocene chalk deposits are discussed in the present chapter). A third unit, the Oxfordian to Barremian Delfland Subgroup of the Schieland Group includes Lower Cretaceous siliciclastic, continental deposits that were laid down locally at the southern, landward side of the Rijnland Group (cf. Wong, this volume). The sequence stratigraphy of the Berriasian to Lower Aptian interval has been dealt with by Hoedemaker & Herngreen (2003). For the Santonian to Danian interval along the southern border of the North Sea Basin reference is made to Vandenberghe et al. (2004).

Schieland Group, Delfland Subgroup

The Schieland Group comprises all mainly continental Upper Jurassic and Lower Cretaceous deposits, including the former Delfland Group (NAM & RGD, 1980) and Delfland Formation (Herngreen & Wong, 1989). The name Delfland Subgroup is in use for the strata of continental origin to the south of the Vlieland High. The Valanginian to Barremian part of this subgroup is dealt with below.

Nieuwerkerk Formation

Van Adrichem Boogaert & Kouwe (1993-1997) divided the Nieuwerkerk Formation in the West Netherlands Basin into the Alblasserdam, Delft Sandstone, and Rodenrijs Claystone members. The Delft Sandstone Member was interpreted as stacked distributary-channel deposits in a lower coastal-plain setting. DeVault & Jeremiah (2002) did not recognize this member as a separate stratigraphic unit

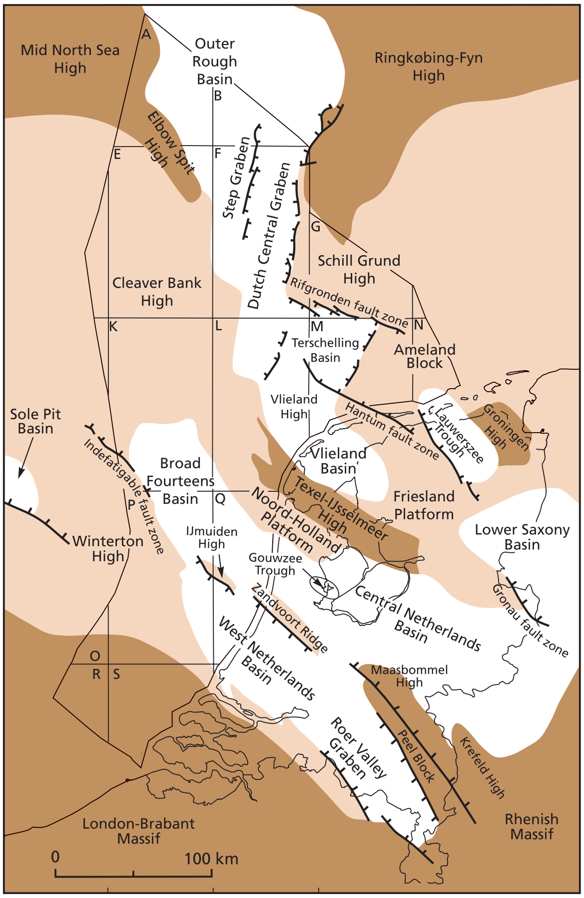

Fig. 2. Map of Late Jurassic to Early Cretaceous structural elements in the Netherlands (after Van Adrichem Boogaert & Kouwe, 1993-1997). Dark brown: structural high, partly subaerial landmass; light brown: platform, intermittently flooded; white: basin.

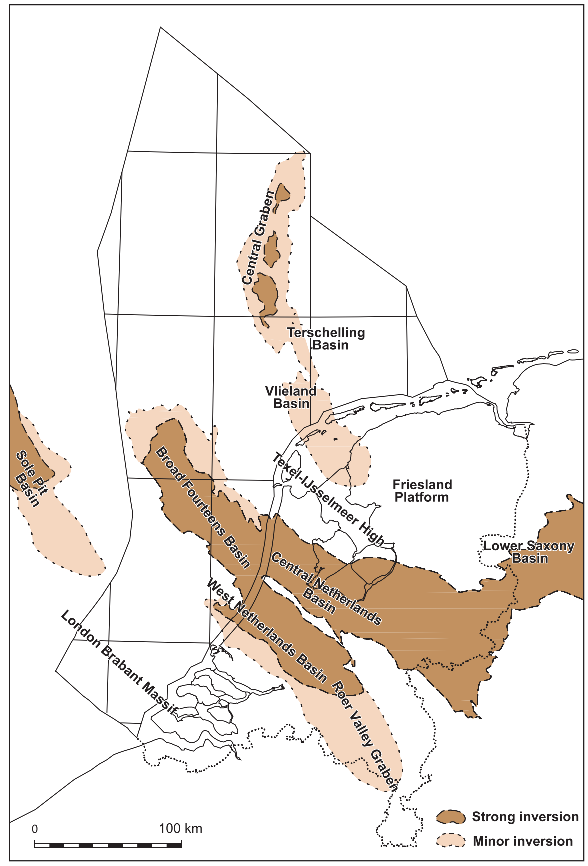

Fig. 3. Inverted basins in the Netherlands. Light shading shows mildly inverted areas, where the Upper Cretaceous chalk is thin as a result of erosion and/or limited deposition; dark shading shows strongly inverted areas, where no chalk has been preserved (De Jager, this volume).

because thick, stacked channel complexes occur throughout the formation. The formation ranges up into the Late Barremian.

The *Alblasserdam Member* is predominantly fluviatile, and red beds are common in overbank settings; the member represents deposition in an upper coastal-plain environment.

The *Rodenrijs Claystone Member*, representing the lower coastal-plain facies, consists of lignitic claystones and lagoonal deposits. The predominance of poorly drained flood-plain deposits, crevasse splays, and minor stacked channel deposits suggests a meandering fluvial system.

Rijnland Group

The largely argillaceous Rijnland Group comprises claystones, marls, siltstones and glauconitic sandstones. It is divided into the Vlieland Sandstone, Vlieland Claystone and Holland formations. The two first-named formations together form the Vlieland Subgroup. Originally, sediments of this group were deposited over the entire Netherlands region, except for the Texel-IJsselmeer High and southern Netherlands. At the locations of inverted, former Jurassic and Early Cretaceous basins, strata belonging to this group have been eroded. The most completely developed section of possibly more than 1400 m is present in the West Netherlands Basin (Fig. 4). The correlation panels of Figures 5 and 6 show log characteristics and lithology of the formations of this group. The depth of its base is highly variable, ranging for example from less than 1000 m in the eastern Netherlands to more than 3000 m in the central and northern offshore region (Fig. 7). Its age ranges from latest Ryazanian or Valanginian to Albian.

Vlieland Sandstone Formation

The Vlieland Sandstone Formation consists of shallow-marine sandstones with subangular to well-rounded, very fine to medium grains; conglomeratic beds occur locally. The sandstones are argillaceous to clean and commonly contain glauconite. Widespread intense bioturbation and wave action have often obscured the primary sedimentary structures. The thickness of the formation varies from

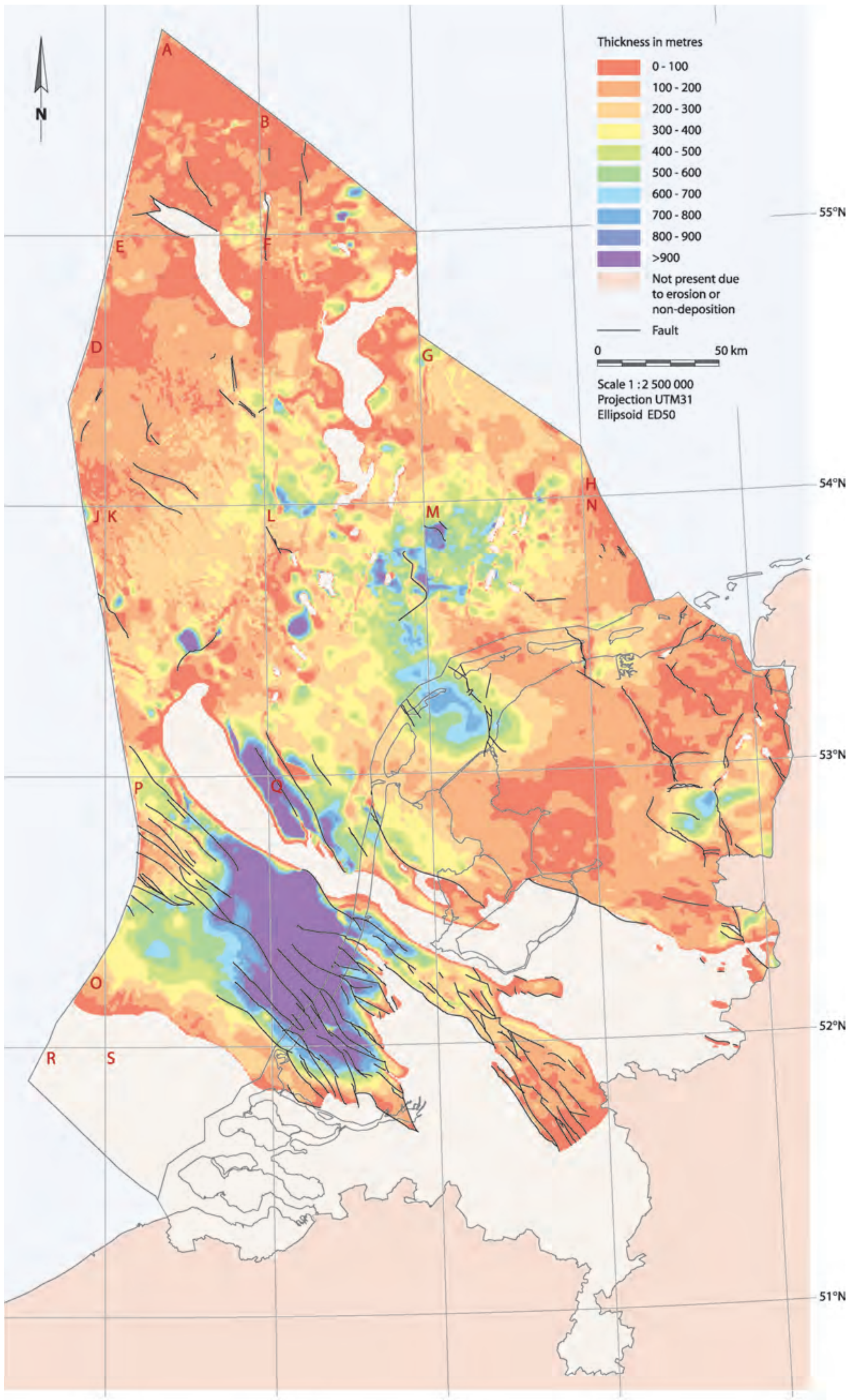

Fig. 4. Isopach map of the Rijnland Group (Duin et al., 2006).

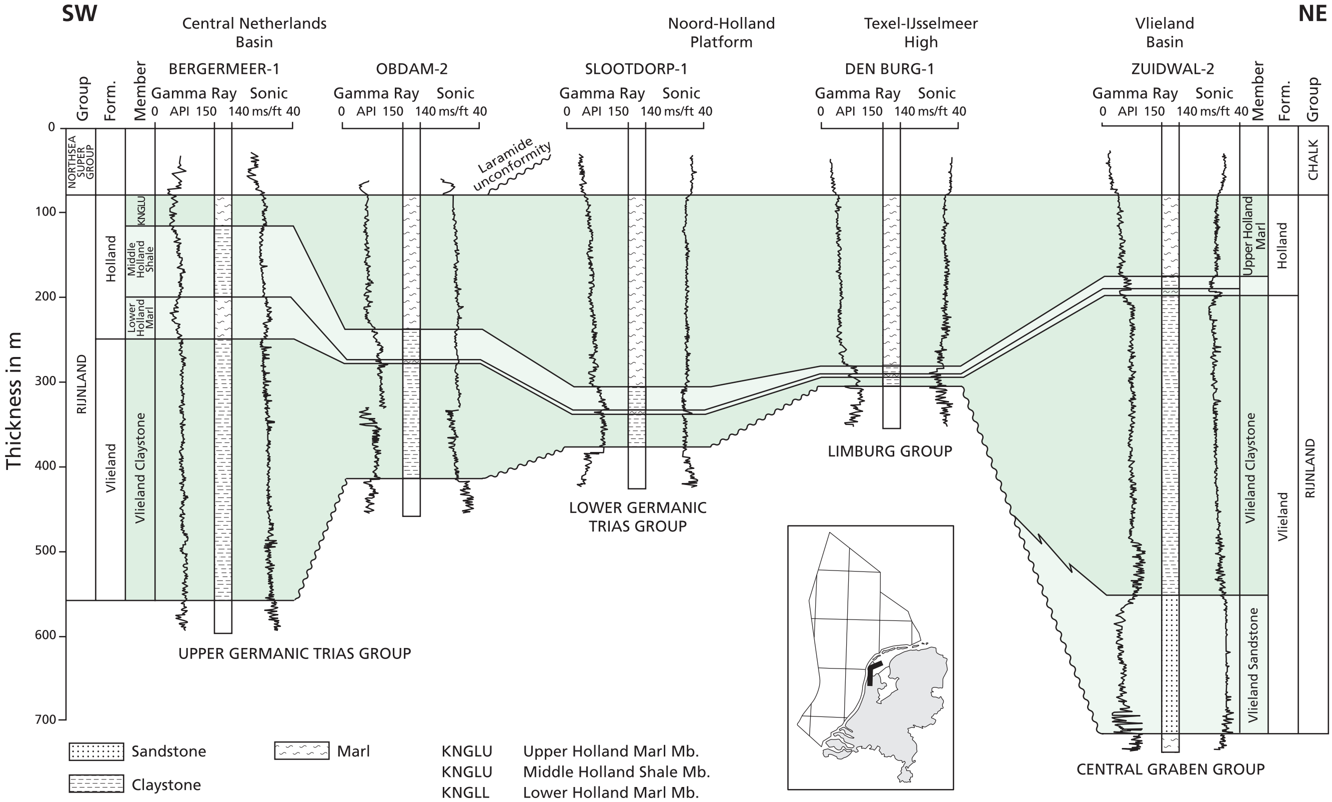

Fig. 5. Well-log correlation (1993 stratigraphy) of the Rijnland Group in the northern Netherlands (after RGD, 1993a). Length of section is ca. 85 km.

basin to basin and reaches a maximum of 200 m in the West Netherlands Basin. Depending on the location, the age of the formation ranges from Ryazanian to Aptian. The formation has been divided into several members many of which are hydrocarbon reservoirs. They have different names per basin (Figs 8a-c):

– northern Netherlands (including Vlieland Basin, southern Dutch Central Graben, Terschelling Basin, Friesland Platform, Groningen High, Noord-Holland Platform/Central Netherlands Basin, Texel-IJsselmeer High): Friesland Member (Cottençon et al., 1975; Perrot & Van der Poel, 1987; RGD, 1991a, b; Herngreen et al., 1991),
– Broad Fourteens Basin: Kotter and Helder members (Roelofsen & De Boer, 1991; De Jong & Laker, 1992; Goh, 1996),
– West Netherlands Basin: De Lier, IJsselmonde Sandstone, Berkel Sandstone, Berkel Sand-Claystone, Rijswijk and Rijn members (Bodenhausen & Ott, 1981; De Jager et al., 1996; Den Hartog Jager, 1996; Racero-Baena & Drake, 1996),
– Lower Saxony Basin: Bentheim Sandstone and Gildehaus Sandstone members (Kemper, 1976, 1992; NITG, 1998, 2000),
– Achterhoek (eastern Gelderland): Kuhfeld Beds (Herngreen et al., 1994, 2000); whether this depositional area should be assigned to the Alstätte Embayment (the southernmost extension of the Lower Saxony Basin: e.g. Kemper, 1976) or to the Central Netherlands Basin as accepted by Burgers & Mulder (1991) and Geluk (in NITG, 1998), remains uncertain.

Vlieland Claystone Formation

The Vlieland Claystone Formation comprises brownish grey to grey claystones with common mica and lignitic matter. Usually they are slightly calcareous; however, in the offshore Dutch Central Graben and Step Graben areas the upper part of the formation is strongly calcareous. In the Broad Fourteens and West Netherlands basins the formation is very silty to sandy with numerous siltstones and very fine sandstone beds. Its thickness is highly variable; a maximum of 350 m is attained in the Vlieland Basin. Its age ranges from Valanginian to Early Aptian and its depositional environment was a shallow to fairly deep-marine (middle to outer-neritic) setting, whereby sand beds were deposited mainly during storms and lowstand periods. The formation has been divided into several members with the following different names per basin (Figs 8a-c):

– northern Dutch Central Graben: Vlieland Marl Member,
– Broad Fourteens and West Netherlands basins: the

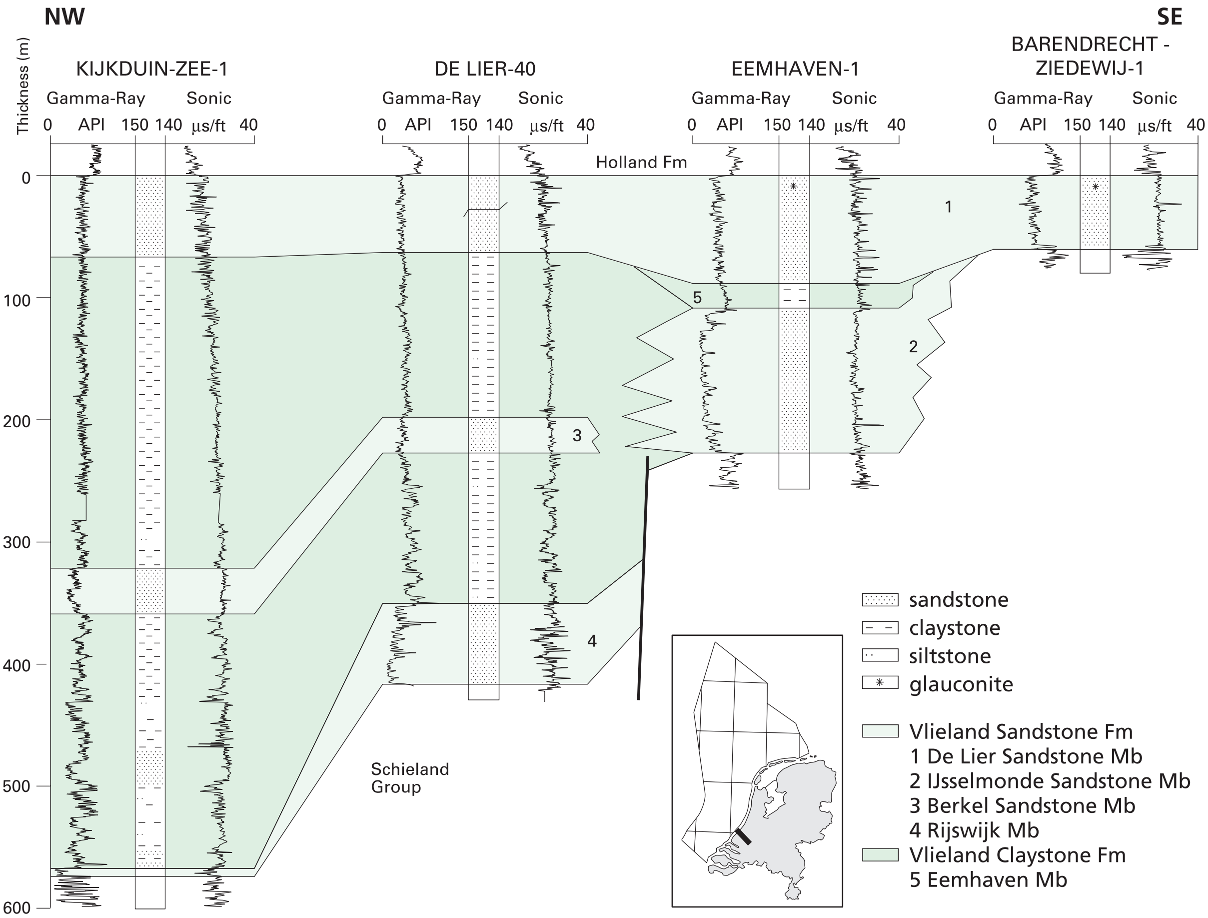

Fig. 6. Well-log correlation of the Rijnland Group in the West Netherlands Basin (after NITG, 2002). Length of section is ca. 35 km.

IJsselmonde Claystone Member (between the Berkel and IJsselmonde Sandstone members) and Eemhaven Member (between the IJsselmonde Sandstone and De Lier members) are recognized in upward direction; towards the basin centre both members grade into the main body of the Vlieland Claystone Formation,

– Lower Saxony Basin: the Bentheim Claystone, Ruinen and Westerbork members; the two last-mentioned members are indistinguishable if the Grenzsandstein is not developed; in that case they are together informally named as Schoonebeek member.

Holland Formation

The Holland Formation consists of grey and reddish brown marls and marly claystones with thick incursions of greensands in the basin-fringe area and thin intercalations of bituminous shales in its lower part. Its distribution covers most of the offshore and northern to central onshore areas. Seismic data, thickness differences and intense reddening suggest the existence of several hiatuses. The unconformable contact with the Vlieland Claystone Formation is highlighted by a distinct upward increase in carbonate content, as reflected by a decrease of gamma-ray readings and by an increase in resistivity and sonic velocity. The total thickness in the northern onshore areas is usually some tens of metres with a considerable increase to more than 300 m in the central parts of the Lower Saxony Basin where in adjacent Germany the formation reaches over 500 m north of Bentheim. In the Broad Fourteens Basin 450 m can be attained. The age of the formation is Aptian-Albian. It was mainly deposited in a fairly deep-marine environment (middle to outer-neritic). Bituminous deposits indicate periods of stagnant basin-floor circulation.

The *Lower Holland Marl Member* consists of grey and variegated (brown, red and yellow) marl or calcareous claystone with intercalated bituminous shales, which are correlated with the German 'Fischschiefer' (e.g. Kemper, 1976). On the Noord-Holland Platform it is developed as limestone. Its thickness in the northern Netherlands is usually ca. 5 to 15 m, but southwest of the Texel-IJsselmeer High, on which it is absent, it reaches more than 60 m. The member is of Aptian to earliest Albian age.

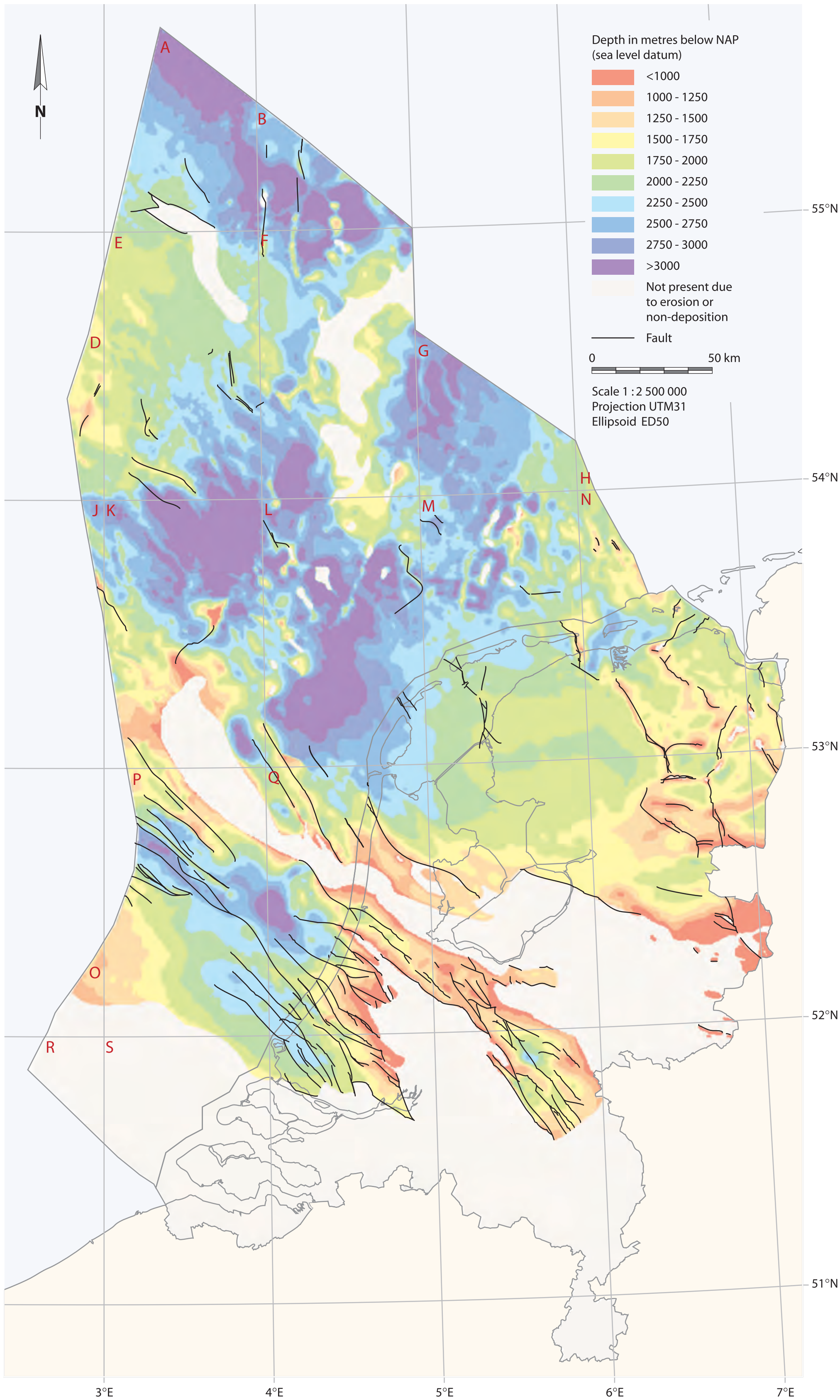

Fig. 7. Depth map of the base of the Rijnland Group (Duin et al., 2006).

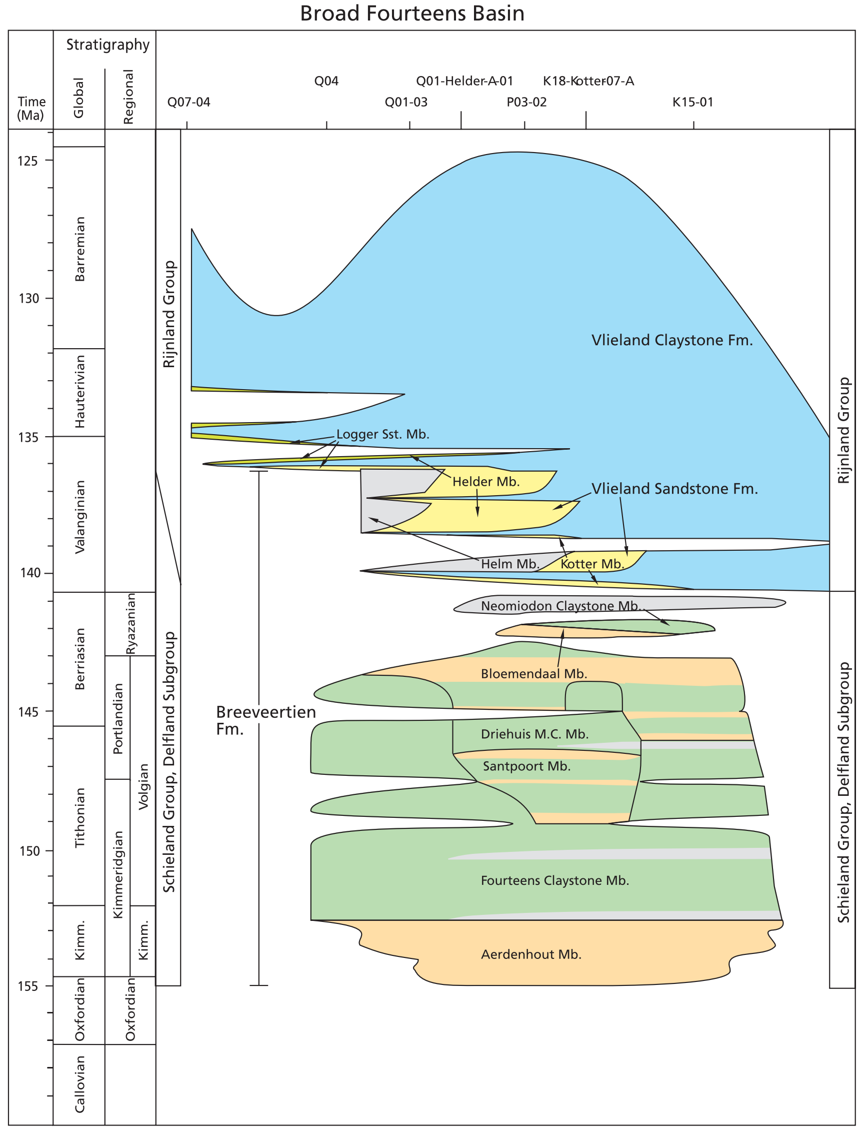

Fig. 8. Stratigraphic schemes of the Valanginian to Barremian Vlieland Sandstone and Vlieland Claystone formations of the Rijnland Group in: a) Broad Fourteens Basin, b) West Netherlands Basin and c) Lower Saxony Basin. After Van Adrichem Boogaert & Kouwe (1993-1997); the codes shown in the legend (d) refer to the nomenclature compiled by these authors.

The *Middle Holland Claystone Member* is a grey and/or reddish brown, calcareous shaly claystone with a distinctly lower calcareous content than the under and overlying members. Unnamed, thin, sandy or conglomeratic beds at its base are included in this member and mark the widespread 'Albian transgression' (Crittenden, 1987). The thickness is often 10 to 20 m in the northeastern Netherlands; in southeast Drenthe and towards the Central Netherlands Basin it increases to up to 100 m. Biostratigraphic data indicate a Late Aptian to Middle Albian age; there are, however, indications in many wells for internal hiatuses or highly condensed sections.

The *Upper Holland Marl Member* consists of light-grey and variegated marls, mainly reddish but also green and violet. It is the thickest of the three members; in the northern Netherlands usually tens of metres with an increase towards the Central Netherlands Basin to between ca. 100 and 200 m. Its age is Middle to Late Albian.

The *Holland Greensand Member* is an alternation of greenish grey, very glauconitic, very fine to fine-grained argillaceous sandstones and locally siltstones. In the Achterhoek area this member contains phosphorite nodules and gravels at its base and rests directly on the Kuhfeld Beds, the hiatus spanning most likely the Barremian.

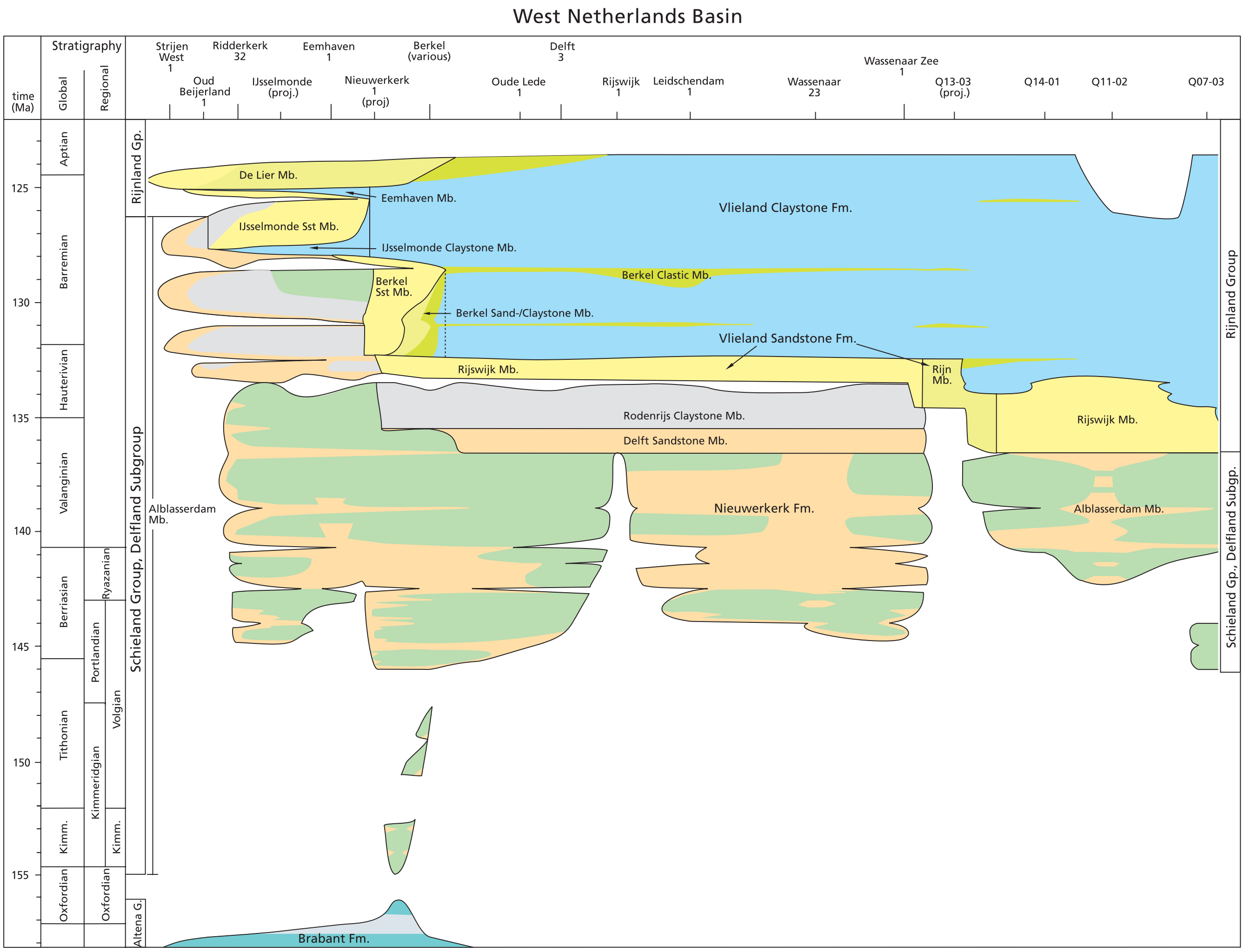

Fig. 8b. (*Continued.*)

Possibly this succession may be compared with the lower part of the Rothenberg Sandstein in Germany (Kemper, 1976). The total thickness of this member in the Achterhoek is at least 80 m (Herngreen et al., 2000). In the West Netherlands Basin and southeastern Broad Fourteens Basin, where it is more than 200 m thick, it grades laterally into a shallow-marine equivalent, the Spijkenisse Greensand Member, along the southern edges of these basins (Fig. 9). The Holland Greensand Member is of Early Albian age, but locally a Late Aptian age has been recorded.

The *Spijkenisse Greensand Member* comprises mainly coarse-grained, glauconitic sandstones, reaching a thickness of 50 to 90 m, still undated.

Chalk Group

The Chalk group consists of a succession of marine, predominantly bioclastic and in part marly limestones (Fig. 10). Flint concretions, isolated or in layers, are common. Locally, other rock types such as glauconitic sandstones occur. The group is present over most of the Netherlands and the offshore area. However, due to erosion resulting from the latest Cretaceous and earliest Tertiary inversion it is locally absent (Fig. 10). In the Dutch Central Graben, for instance, the group is missing, or present as a patchy, thin cover. The group's maximum thickness is more than 1800 m (Fig. 11). Figure 12 shows log characteristics of the formations of the group in the north of the country. The depth of the base of the group is more than 2750 m in the northern and central-western offshore area (Fig. 13).

The Chalk Group in most of the Netherlands is divided (from bottom to top) into the Texel, Ommelanden and Ekofisk formations. The time-stratigraphic equivalents of the Ommelanden Chalk as exposed in southern Limburg or drilled on the Peel Block are assigned to the siliciclastic Aken, Oploo and Vaals, and the calcareous Gulpen and Maastricht formations. The Ekofisk Formation, the Houthem Formation in Limburg and the uppermost part of the Maastricht Formation are of Danian age and represent the oldest Tertiary strata. The

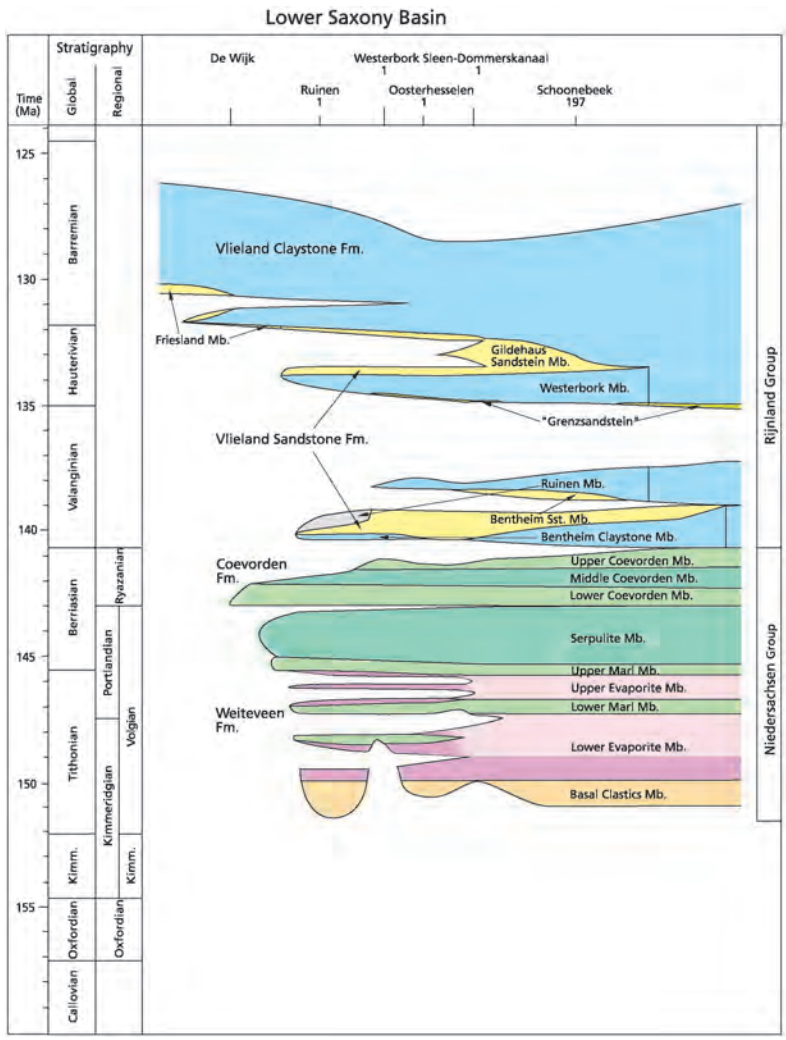

Fig. 8c. (*Continued*.)

age of the Chalk Group ranges from Cenomanian to Danian.

Texel Formation

This formation comprises usually light-grey to beige and white limestones and marly chalks with some marl intercalations. Near the southern edge of its distribution, i.e. the southern part of the West Netherlands Basin and along the Maasbommel High, greensands form its lower part. Elsewhere, in the central part of the West Netherlands Basin, and in the Central Netherlands and Broad Fourteens basins, the formation becomes increasingly marly and in a still more offshore setting north of these basins, it develops into pure limestones.

The formation's thickness varies from usually ca. 50 to 70 m in the northern and eastern Netherlands (Vlieland Basin, Friesland Platform, Groningen High and eastern Overijssel) and increases southwards to 200 m (northern edge of Central Netherlands Basin). The formation is

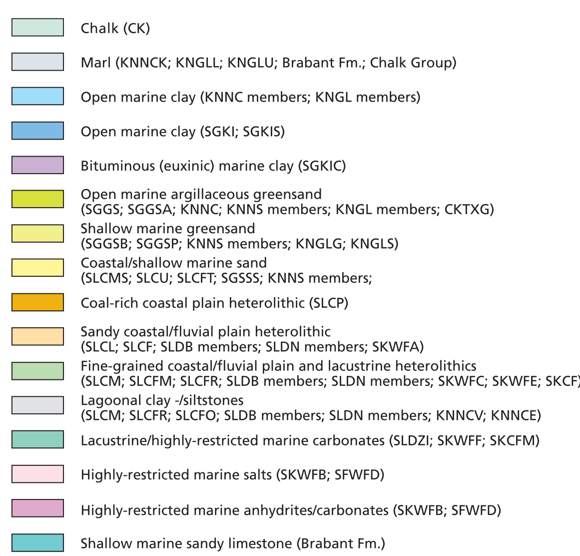

Fig. 8d. Legend to Figures 8 to 10.

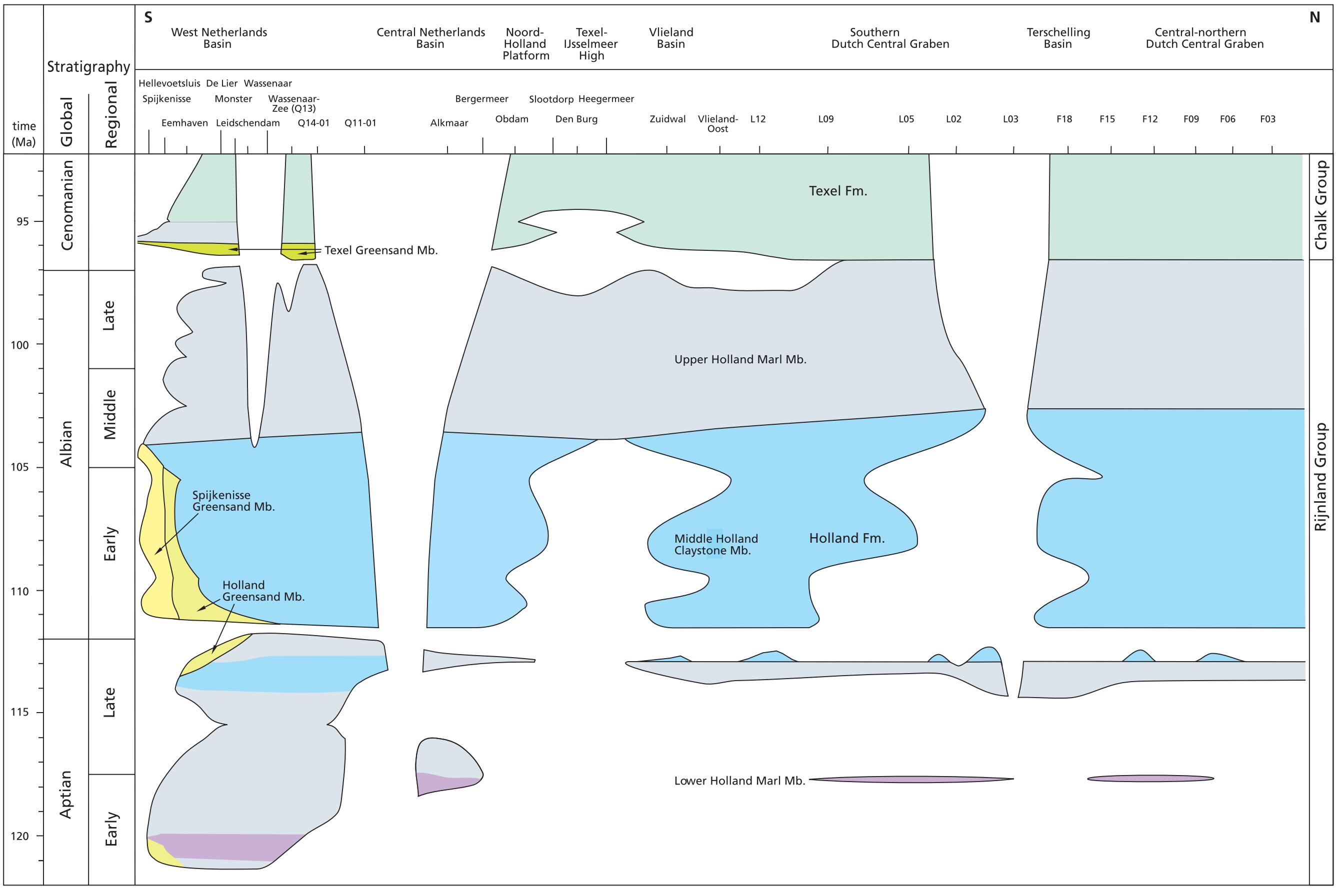

Fig. 9. Stratigraphic scheme of the Aptian-Albian Holland Formation of the Rijnland Group (after Van Adrichem Boogaert & Kouwe, 1993-1997). Legend in Fig. 8d.

locally exposed in the Achterhoek and well developed in adjacent Germany. Its age is Cenomanian to earliest Turonian.

The *Texel Greensand Member* is a highly glauconitic, calcareous sandstone with intercalated marls. It occurs only along the southern fringe of the West Netherlands Basin, approximately from Rotterdam eastwards to Nijmegen.

The *Texel Marlstone Member* consists of white to light-grey limestones and marly chalks in the northern Netherlands and of marls with marly chalks south of the Broad Fourteens Basin and in the western part of the West Netherlands Basin. Like the Texel Greensand, its distribution is bounded to the south by a line running from Rotterdam to Nijmegen.

The *Plenus Marl Member* is a dark-grey, locally black, calcareous, laminated, bituminous claystone of widespread distribution. Its thickness is usually some decimetres to a few metres and it is easily recognized on wireline logs by the characteristic high(er) gamma-ray and resistivity, and lower sonic readings (Fig. 12). It is of latest Cenomanian age.

Ommelanden Formation

This formation is a succession of white to beige or light-olive grey, fine-grained, and in places argillaceous limestones. Layers of flint nodules are locally common in its Campanian to Maastrichtian ('Upper Senonian') part. The Turonian interval is a whitish, hard, dense limestone with high sonic readings. It is succeeded by a more marly succession of Coniacian to Santonian age (informally called middle Ommelanden; NITG, 2001). This, in turn, is overlain by a less marly section (upper Ommelanden) which, in its lower part, consists of consolidated calcarenites, grading upwards into massive chalk with abundant flint nodules.

Biostratigraphic and wireline-log correlation studies indicate several intraformational unconformities (Herngreen et al., 1996). Along the basin edge in the southeast Netherlands, tongues of sandstones, greensands and in part coarse, bioclastic limestones occur. In the northern Netherlands the maximum thickness is usually just over 1000 m. The regional differences here are due to salt movements in the subsurface. Thicknesses up to 1500 m are reached in the offshore K and L quadrants. Due to erosion following inversion phases, the formation is locally very thin or even absent. It was deposited under rela-

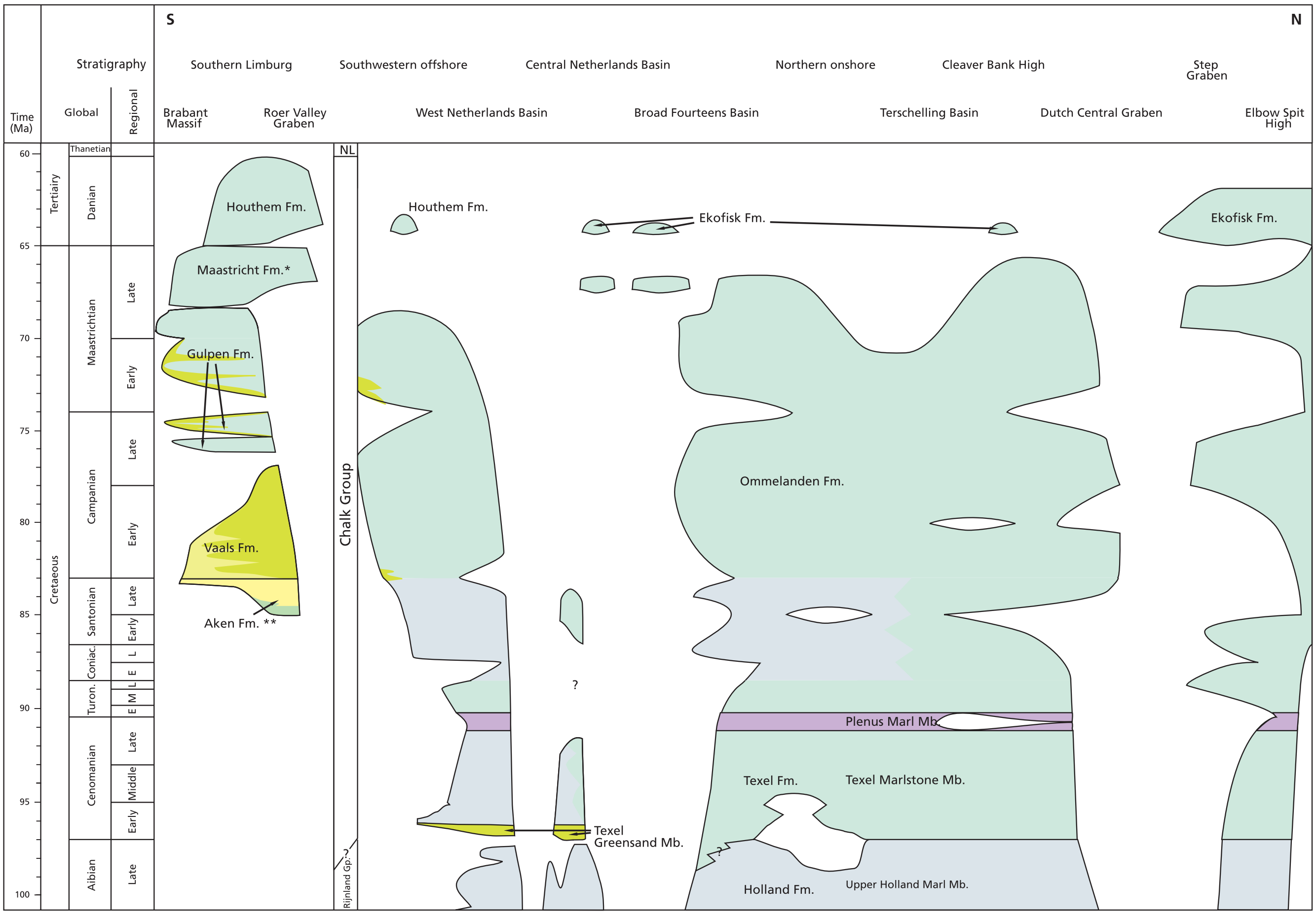

Fig. 10. Stratigraphic scheme of the Cenomanian-Danian Chalk Group (after Van Adrichem Boogaert & Kouwe, 1993-1997). *) The uppermost part of the Maastricht Formation is Danian. **) Oploo Formation in Peel Block area. Legend in Fig. 8d. NL: Lower North Sea Group.

tively stable, low-energy conditions in carbonate-shelf and upper-bathyal settings.

Ekofisk Formation

This formation comprises white, chalky limestones with rare nodular and bedded flint layers, and thin, grey to green clay laminae; glauconite may occur in its basal part. The formation displays a characteristic gradual upward increase in acoustic and resistivity-log readings. Rarely a gamma-ray peak is seen at its base, representing the Cretaceous/Tertiary boundary clay, which equates with the Fish Clay in Denmark (e.g. Schmitz et al., 1992) and the clay in unit IVf-7 (upper part of Meerssen Member) in southern Limburg (Smit & Brinkhuis, 1996; Herngreen et al., 1998). Near this boundary a hardground, overlain by a glauconite-rich bed, occurs locally. The formation is found in the northern and southern parts of the Dutch North Sea sector and in parts of the West and Central Netherlands basins. Its thickness attains maximum values of 100 m. It is of Danian age and was mainly deposited under similar conditions as the Ommelanden Formation but locally, near salt domes, redeposition by gravitational mass flow occurred.

The Chalk Group in the southeast Netherlands

The Chalk Group in southern Limburg rests unconformably on Paleozoic, and to the north also on Triassic and Jurassic rocks. Felder (1975) and Felder & Bosch (2000) formally described its lithostratigraphic units, which are of Santonian to Danian age. They distinguished five formations (Fig. 10), each consisting of several members which are usually separated from each other by regionally recognizable hardground horizons or omission surfaces. The chalk in the area yielded several large mosasaurs (e.g. Dortangs et al., 2002).

Aken Formation This is a succession of well-sorted, fine-grained quartz sands and smectitic clays. In the clays, root horizons, in part associated with thin lignite layers, are common. The formation is restricted to the southern parts

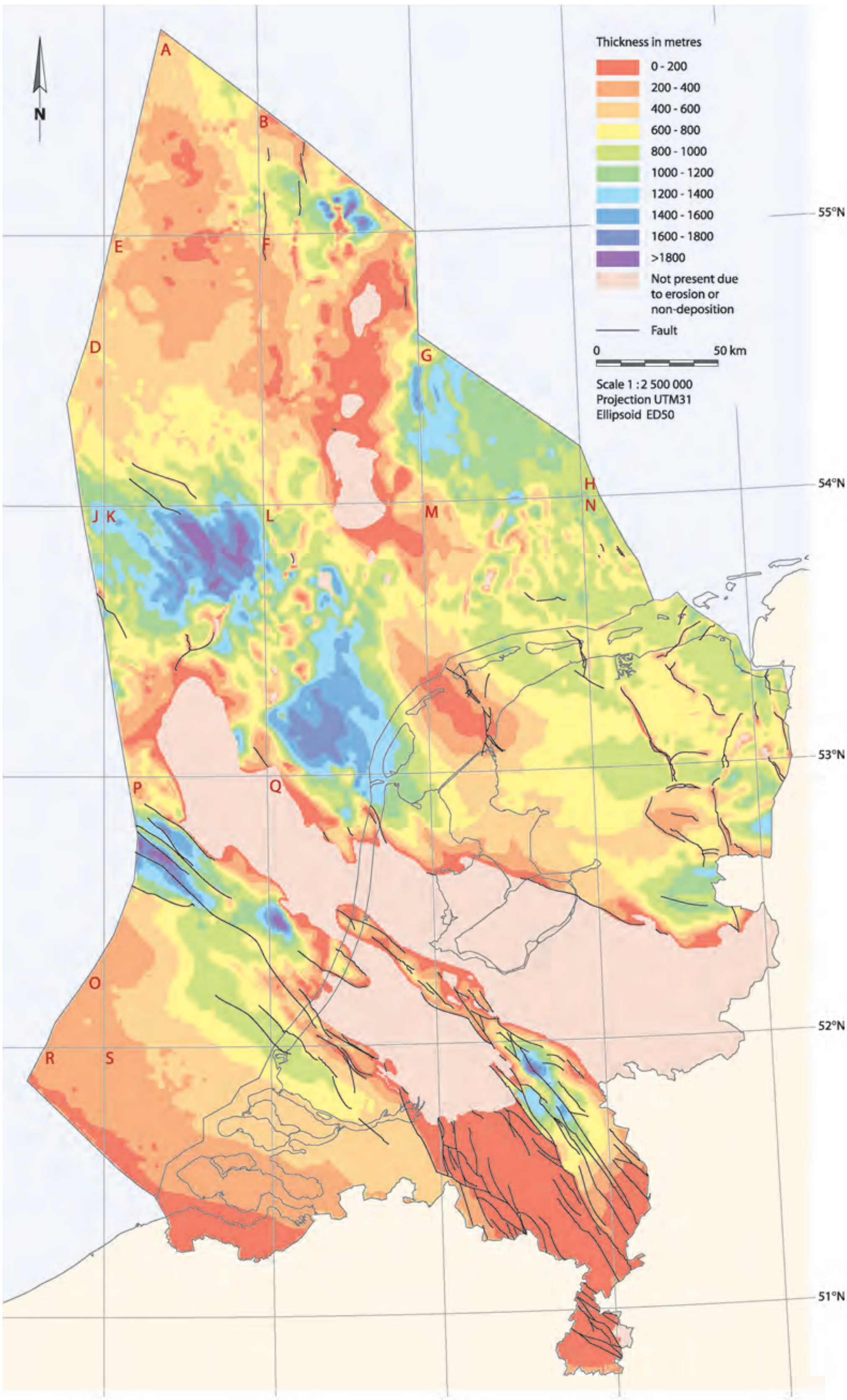

Fig. 11. Isopach map of the Chalk Group (Duin et al., 2006).

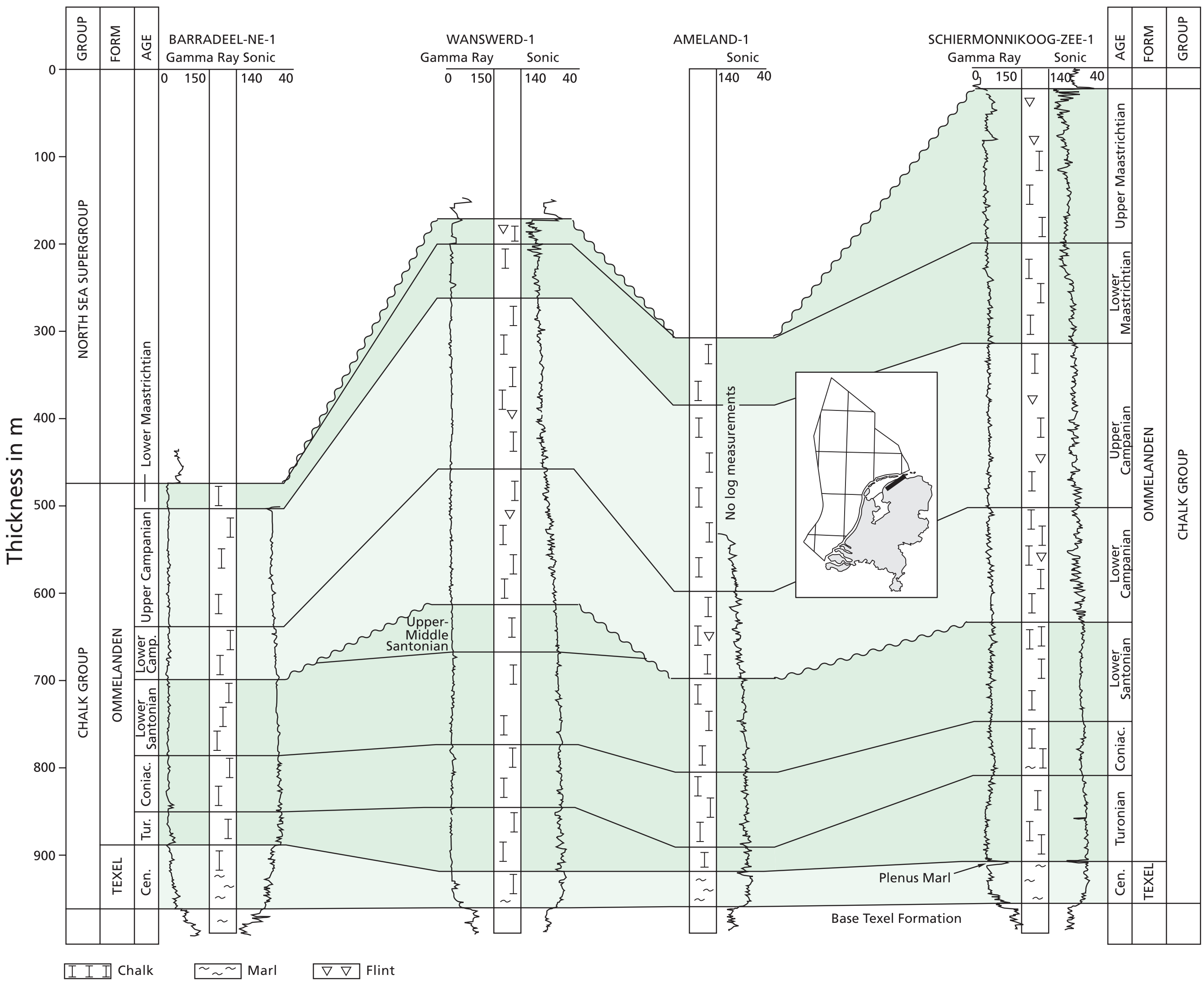

Fig. 12. Well-log correlation of the Chalk Group in the northern Netherlands (after RGD, 1991b). Length of section is ca. 55 km.

of the provinces Noord-Brabant and Limburg where its thickness varies from less than 1 to ca. 60 m. In the Peel Block area its lateral equivalent is the Oploo Formation (Gras & Geluk, 1999; NITG, 2001). The formation, Santonian of age, is divided into three members: Hergenrath Clay, Aken Sand and Hauset Sand. It was deposited under fluvial to shallow-marine conditions. (Aken: Dutch name for Aachen, Germany.)

Oploo Formation For the Peel Block area, Gras & Geluk (1999) introduced the Oploo Formation. This formation is a coarsening-upward, partly glauconitic, siliciclastic succession. It can be correlated with the Emscher Marl and Haltern-Recklinghausen Sands in adjacent Germany. Towards the north, the sandstones of the Oploo Formation grade into marls and chalks of the Ommelanden Formation. The formation represents a marine basin-edge facies and is more than 200 m thick. It is the lateral equivalent of the Aken Formation in southern Limburg and adjacent areas. It is of Santonian age.

Vaals Formation The Vaals Formation consists in its type area in southern Limburg predominantly of decalcified, yellow to greyish green, fine-grained, glauconitic sand bodies, each of which has a silty or clayey base; their top parts are usually bioturbated. To the west and northwest, the formation passes into calcareous, glauconitic silty clay. Locally it is highly fossiliferous. Masses of marine shells and glauconite grains are concentrated in channel fills. The Vaals succession in southern Limburg starts with a 25 to 50-cm-thick basal conglomerate on top of the bioturbated Hauset Sand (Aken Fm). Seven members have been described; for details reference is made to Felder & Bosch (2000). Their mainly shallow-marine deposition was in-

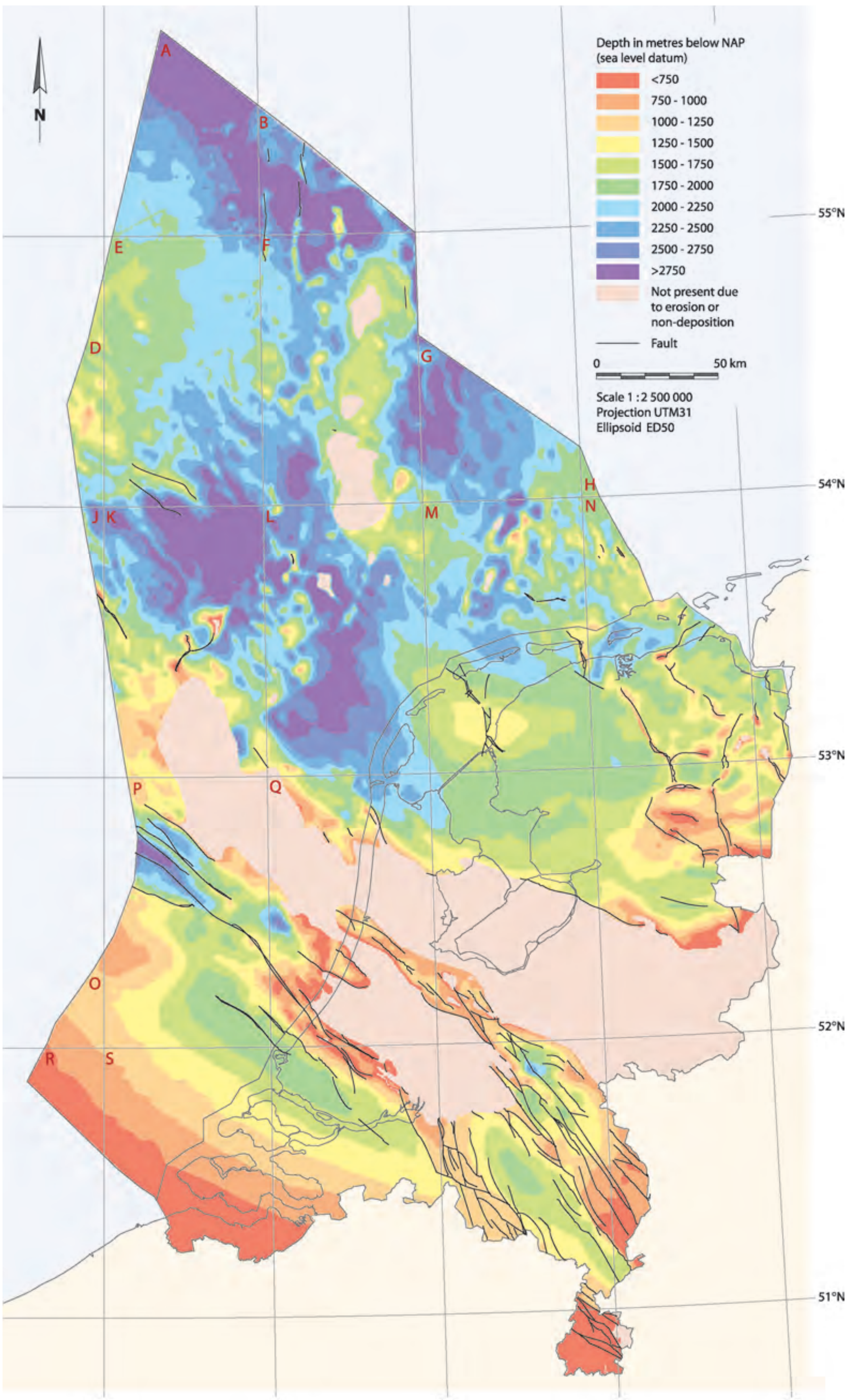

Fig. 13. Depth map of the base of the Chalk Group (Duin et al., 2006).

terrupted by tectonic movements which caused submarine erosion of major parts of the formation. The distribution of the formation is essentially restricted to areas in Noord-Brabant and Limburg. Due to the synsedimentary movements and post-depositional erosion the thickness varies from ca. 1 to 200 m. The formation is of early Early to early Late Campanian age; thus, the higher portion is the lateral time-equivalent of the Zeven Wegen Member (Gulpen Fm) (Jagt, 1999).

Gulpen Formation This formation consists predominantly of soft, fine-grained limestone with varying amounts of glauconite and flint nodules. It was deposited under fully marine conditions. Its distribution is restricted to areas in Noord-Brabant and Limburg, and its total thickness varies from 1 to ca. 175 m. Details regarding distribution, thickness, lithology, exposures etc. of each member are treated by Felder & Bosch (2000). The formation is of Late Campanian to early Late Maastrichtian age (Herngreen, 1998; Jagt, 1999). It includes an unconformity at the base of the Vijlen Member, which in southern Limburg locally lies on Paleozoic rocks.

In the Eastern Campine area and the westerly Roer Valley Graben border zone in Belgium, Felder et al. (1985) distinguished a transitional facies of the Gulpen Formation, provisionally indicated as the 'Pre-Valkenburg Strata', which is characterized by glauconitic marls. A three-fold subdivision can be recognized on the basis of higher gamma-ray readings with a peak in the middle part, which corresponds to the lower part of the Beutenaken Member (provisionally indicated as Beutenaken Marl). NITG (2001) mentioned a similar development near the Dutch part of the inverted Roer Valley Graben and on the Peel Block.

Maastricht Formation Two facies types are distinguished within the Maastricht Formation: in the southwestern part of southern Limburg the Maastricht Limestone, soft, yellow-whitish with 5 to 12% flint, and in the northeast the Kunrade Limestone, an alternation of hard and soft, light-grey layers without silica concretions. The limestone of Schaelsberg forms an intermediate development (Felder & Bosch, 2000). Cryptocrystalline quartz concretions (flint or chert) of different sizes, morphology and abundance, are mainly concentrated in layers but also occur isolated. The thickness of the formation ranges from 45 to 90 m. Its sedimentary environment was marine. Its age is Late Maastrichtian and Danian. The Maastrichtian in the Maastricht area was first described by Dumont (1849) as a separate unit resting on top of the 'système sénonien'. The quarry of the ENCI cement factory is its former type locality.

The Valkenburg, Gronsveld, Schiepersberg and Emael members and the Kunrade Limestone display a distinct cyclic development. The overlying member, the Nekum Limestone, is a coarse-grained chalk with poorly developed cyclicity and rare flint concretions. The concretions and their origin were thoroughly studied by Zijlstra (1994). The youngest member, the Meerssen Limestone, is characterized by numerous detrital fossil layers of algae, bryozoans, corals, etc. above hardgrounds, and does not contain flint. Its uppermost part, IVf-7, is of Danian age (Brinkhuis & Smit, 1996).

Houthem Formation This unit consists of soft, light-grey to beige, fine to coarse-grained limestones with intercalated hard limestone bands and indurated limestone concretions. The lowermost part of the formation is characterized by the occurrence of glauconite. The formation's occurrence is restricted to parts of Noord-Brabant and Limburg. The thickness varies from 1 to over 50 m.

Its lower boundary is the Vroenhoven Horizon, a conspicuous hardground with bioturbations at the top of the Maastricht Formation. The Houthem Formation is subdivided in ascending order into the Geulhem Limestone, exposed at Curfs quarry (Herngreen et al., 1998; Felder & Bosch, 2000), and the Bunde and Geleen limestones. The latter two are only known from boreholes and mine shafts. They consist of coarse-grained, yellowish white limestone with hard bands and of very coarse-grained, probably cyclic, limestone. The formation is of Danian age and represents sedimentation in a shallow-marine, carbonate environment.

Geological history

In the Early Cretaceous, the differential basin movements of the Jurassic were succeeded by a period of relative tectonic quiescence. Isostatic compensation following Jurassic tectonic uplift led to regional subsidence of the post-rift sag phase. Independently of this, a new eustatic transgressive phase began in the Valanginian. This phase is related to the opening of the southern Atlantic Ocean and the Indian Ocean (cf. De Jager, this volume).

As mentioned by Wong (this volume), differential subsidence in the Dutch Central Graben terminated with the Late Kimmerian II pulse (Late Ryazanian). A more evenly distributed sedimentation pattern, associated with the post-rift phase, began. The general overstepping nature of the basal Cretaceous sediments onto existing basins and highs, and the consecutive establishment of extensive uniform post-rift sedimentation during the later Cretaceous can be clearly seen on seismic sections (Fig. 14). For the Late Ryazanian to Early Valanginian interval Hoedemaeker & Herngreen (2003) distinguished three Type 1 sequence boundaries.

The *Valanginian* transgressions simultaneously reached the Vlieland, Broad Fourteens and Lower Saxony basins (Fig. 15), where the first transgressive and coastal-barrier

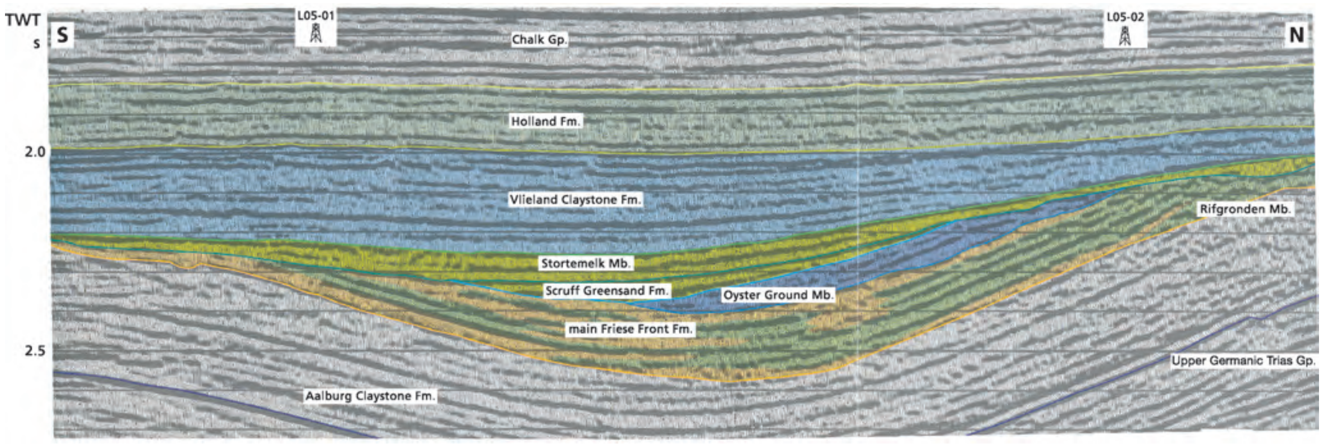

Fig. 14. Seismic section of southernmost part of Dutch Central Graben (block L5) showing overstepping by the Stortemelk Member (Scruff Group) of a former Jurassic basin and the establishment of extensive uniform sedimentation of the Vlieland Claystone and Holland formations during the Early Cretaceous (after Van Adrichem Boogaert & Kouwe, 1993-1997). In this section, base Stortemelk represents the Late Kimmerian Unconformity. Length of section is ca. 17 km.

sands of the Vlieland Sandstone Formation (Friesland, Kotter and Bentheim Sandstone members, Figs 8a-c) were deposited. Towards the east and north of the Broad Fourteens Basin, the transgressive Kotter Member laterally changed into the lagoonal Helm Member of the Breeveertien Formation (Fig. 8a), while the Friesland Member in the Vlieland Basin was followed by open-marine muds of the Vlieland Claystone Formation. In the Lower Saxony Basin the open-marine Bentheim Claystone Member locally preceded deposition of the Bentheim Sandstone Member (Fig. 8c). The latter was followed by the open-marine to lagoonal Ruinen Member. During the Valanginian, the role of the Vlieland Basin as depocentre ended. The coastline shifted to the Friesland Platform and the central parts of the Broad Fourteens Basin, where stacked coastal-bar complexes were formed (Helder Mbr). In the West Netherlands Basin the fluvial 'Delft Sandstone Member' and the lagoonal Rodenrijs Claystone Member of the Nieuwerkerk Formation are the coastal and fluvial-plain equivalents of the coastal-barrier complex in the Broad Fourteens Basin. Inland, fluvial-plain deposition occurred in the Roer Valley Graben and southern West Netherlands Basin (Van Adrichem Boogaert & Kouwe, 1993-1997; DeVault & Jeremiah, 2002) with the Alblasserdam Member of the Nieuwerkerk Formation, and also in the Central Netherlands Basin and on the Noord-Holland Platform (Zurich Fm). In the Dutch part of the Lower Saxony Basin, the Late Valanginian and Early Hauterivian history is obscured by a hiatus, separating the Ruinen and Westerbork members (Fig. 8c). This hiatus is of regional extent since it is also known from the Vlieland Basin (Herngreen et al., 1991) and West Netherlands Basin (Haanstra, 1963). It belongs to the complex of unconformities at or near base Rijnland which is often referred to as 'Late Kimmerian Unconformity' and which more or less

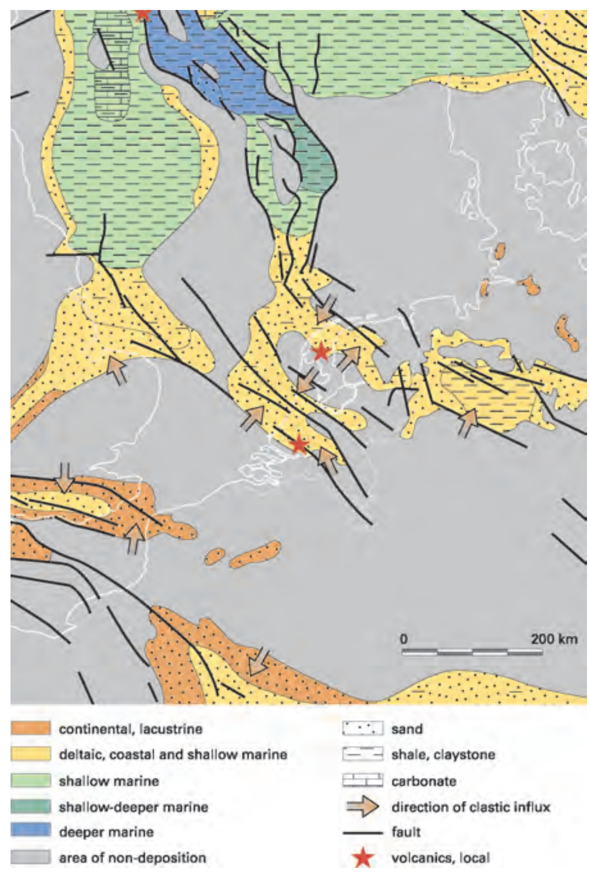

Fig. 15. Paleogeographic map of the Netherlands and adjacent areas during the Ryazanian-Valanginian (after Ziegler, 1990).

marks the transition from syn-rift to post-rift deposition.

During the *mid-Hauterivian* transgressive phase, the sea covered an increasing portion of the West Netherlands and Central Netherlands basins and various platform areas. The coastline shifted from the Broad Fourteens Basin to the West Netherlands and Central Netherlands basins, Noord-Holland Platform, southern Friesland Platform and Groningen High, where an amalgamated sheet of transgressive and coastal-barrier sands was deposited (Rijn, Rijswijk and Friesland members of Vlieland Sandstone Fm) followed by marine muds (Vlieland Claystone Fm). On the landward side of the coastal-barrier system of the Vlieland Sandstone Formation in the West Netherlands Basin and Roer Valley Graben, coastal or fluvial-plain and lagoonal deposition of the Delfland Subgroup continued (Nieuwerkerk Fm, former Delft Sandstone Mbr and Rodenrijs Claystone Mbr, Fig. 8b). In the Lower Saxony Basin, intercalated sandstones of mid-Hauterivian age (Gildehaus Sandstone Mbr, Fig. 8c) reflect a period of coastal influence. In the more distal basin areas (Cleaver Bank High, Mid North Sea High), chalk-like marl deposition set in (Vlieland Marl Mbr). The connection between the Vlieland and Lower Saxony basins was established during the Late Hauterivian when a corridor between the Texel-IJsselmeer High in the southwest and the Groningen High in the northeast came into existence (RGD, 1993a, b; Rijkers & Geluk, 1996).

During the *Barremian*, three transgressive-regressive sequences were deposited in the West Netherlands Basin. The first one formed the Berkel Sand-Claystone and Berkel Sandstone members (Fig. 8b). During the second transgressive pulse, the IJsselmonde Claystone and Sandstone members were deposited, and the transgression reached the Rotterdam area. The coastal and fluvial-plain conditions of the Nieuwerkerk Formation retreated further south. At the same time, the crest of the Texel-IJsselmeer High was gradually flooded. The third transgressive pulse included the entire West Netherlands Basin as the Eemhaven Member was deposited. During the subsequent regressive phase in the Late Barremian-Early Aptian, the progradational De Lier Member was laid down along the southern margin of this basin.

In the *Aptian*, the sea had flooded all former basins and adjacent platforms, and open-marine conditions were established over a wide area of the northwest European siliciclastic shelves. The resulting sedimentation of marls and greensands of the Holland Formation was interrupted several times by tectonic activity of the Austrian phase, causing intra-formational hiatuses and variegated marls. A major hiatus occurs near the Aptian/Albian boundary in the northern and eastern Netherlands. In the more rapidly subsiding Central and West Netherlands basins, in contrast, sedimentation probably was fairly continuous.

The fining-upward succession of the preceding Vlieland Claystone culminates in a 'black shale' during the peak transgression in the Early Aptian (Jacquin et al., 1998). Such organic-rich levels are fairly common in the Dutch pre-Albian Cretaceous, i.e. mainly in the Early Barremian part of the Vlieland Claystone, where they are lignitic, and in the middle Early Aptian part of the Lower Holland Marl. They were deposited under anoxic conditions. Contemporaneous 'black shales' are present in the German part of the Lower Saxony Basin, the Hauptblätterton and Fischschiefer respectively (Mutterlose & Bornemann, 2000); like in the Netherlands, both were followed by a Type 1 sequence boundary (Hoedemaeker & Herngreen, 2003). These widespread shales mirror transgressive developments which are not local, as previously stated in the literature, but at least regional and possibly worldwide. The Early Aptian event, for instance, may be correlated with the Oceanic Anoxic Event IA (Kauffman & Hart, 1996; Fig. 16). Sediments of latest Aptian to Early (? and earliest Middle) Albian age are poorly documented in the Netherlands due to erosion related to the Austrian tectonic phase and lack of reliable samples. The significance of the Austrian phase and the subsequent Albian transgression in the Netherlands is, in the present authors' opinion, underestimated. They form a well-documented (Crittenden, 1987) regional event in England, and comparisons

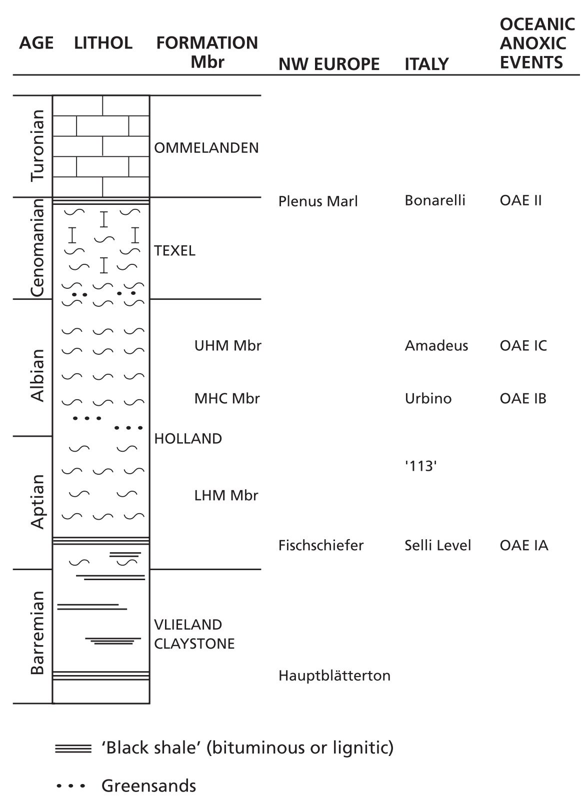

Fig. 16. Correlation of Cretaceous 'black shales' in the Netherlands (with simplified lithology and lithostratigraphy) with those from northwest Europe, Italy and the Oceanic Anoxic Events (based on TNO data; Kauffman & Hart, 1996; Mutterlose & Bornemann, 2000).

with sediments of the contemporaneous Holland Formation indicate that the existing lithostratigraphic model for this formation is a simplification that deserves further study.

A major sea-level rise led to the *Albian* transgression (Crittenden, 1987), and fault activity ceased completely, at least in the area of the Texel-IJsselmeer High and Friesland Platform. The Albian coastline was situated approximately along the line running west to east from southern Zuid-Holland (Rotterdam) to southern Gelderland (Nijmegen). The Texel-IJsselmeer, Noord-Holland and North Netherlands highs were completely inundated in Middle Albian time (the North Netherlands High consists of the Friesland Platform and Groningen High together); on these highs several unconformities and disconformities are present in the Holland Formation. The Austrian phase reactivated various tectonic highs in the northeastern Netherlands (NITG, 1998) and caused a considerable hiatus. Locally, even the entire Rijnland Group was eroded. Around the Albian/Cenomanian boundary, the influx of fine-grained clastics into the marine realm diminished because the adjoining source areas became submerged during the widespread transgression, and the carbonate content, mainly in the form of remains of planktonic organisms, notably coccolithophorid algae, which settled from suspension, increased.

The *Cenomanian* transgression in northwest Europe coincided with basin subsidence which resulted in a rapid increase in accommodation space (Hardenbol & Robaszynski, 1998). During the Early Cenomanian, the Dutch part of the coastline ran west to east from Rotterdam to Nijmegen. At the basin margin, shallow-marine conditions prevailed which are reflected by greensands (Texel Greensand Mbr) and thick, marly deposits (Texel Marl Fm). To the north, these marls grade into pure chalk. Of the six Cenomanian sequences identified in the Anglo-Paris Basin by Robaszynski et al. (1998), only one can be demonstrated with certainty in the Netherlands, viz. the Late Cenomanian sequence Ce5 (Chart 4 of Hardenbol et al., 1998). It corresponds with the Plenus Marl Member which shows a striking increase in clay content. Anoxic conditions led to deposition of this thin, dark-coloured and bituminous marl bed. This event is correlated with the short-term Oceanic Anoxic Event II (Fig. 16) which is recognized worldwide; both (bio-)events are near a latest Cenomanian/earliest Turonian major transgressive/regressive boundary.

During *Turonian to Maastrichtian* times, marine carbonate deposition of the Ommelanden Formation prevailed, forming the bulk of the Chalk Group. Influx of terrigenous material caused deposition of marly intercalations, the distribution of which seems to be related to areas where inversion occurred. The coastline gradually moved southwards and by Campanian times even the London-Brabant Massif and the Rhenish Massif (S quadrant offshore and southern Limburg) were inundated, like in the north the Mid North Sea and Rynkøbing-Fyn highs.

The basins with their Late Jurassic and Early Cretaceous fills underwent inversion in several pulses during the Santonian-Campanian (Herngreen et al., 1996) Subhercynian tectonic phase (Van Wijhe, 1987a, b; Geluk et al., 1994; Nalpas et al., 1996; Gras & Geluk, 1999; De Jager, this volume). Older, in part Paleozoic, WNW-ESE trending faults were reactivated, and those which acted as normal faults during the Kimmerian rifting, behaved in Late Cretaceous times as reverse faults. The depocentres of the basins were uplifted, and consequently deeply eroded. In contrast, a thick succession of chalk covers areas that were not inverted, such as the former highs and platforms, which were downwarped towards the end of the Cretaceous. At that time, the Upper Cretaceous in the northern Netherlands was slightly folded along NW-SE striking axes.

The Santonian to Maastrichtian formations in southern Limburg reflect transgressions over the northern flank of the Brabant Massif. Their deposition started with fluviatile to limnic clays with rootlet soils and sometimes peat layers (Hergenrath Clay), and bioturbated, small-scale cyclic, strongly tidally influenced quartz sands (Aken Sand), followed by laminated, large-scale cyclic sands deposited in a somewhat deeper, marine environment (Hauset Sand). They became covered by glauconite-rich, fine-grained sands alternating with silty and clayey sand in a probably middle to outer-neritic setting (Vaals Fm). In the Late Campanian, the influx of siliciclastics decreased and coccolith-rich lime-muds, with some glauconite in their lower part, were deposited. The silica in these muds became later concentrated as layers or nodules of flint in specific intervals of the Gulpen and Maastricht formations. Hardgrounds and layers of coarse-grained fossil fragments are interbedded in both formations as well, and various unconformities indicate erosion or non-deposition as a result of Late Cretaceous tectonic movements (Schiøler et al., 1997). The succession described by Gras & Geluk (1999) from the Peel Block shows a similar development. In this area, two factors played a prominent role. Next to the Late Cretaceous continuation of the transgression, the Subhercynian inversion of the Roer Valley Graben (Geluk et al., 1994) and the Central Netherlands Basin caused considerable erosion and the now subsiding former highs of the Peel and Campine blocks received this eroded material.

Chalk deposition continued during the Danian (Early Paleocene) with the Ekofisk Formation in carbonate-shelf and upper-bathyal environments (Fig. 17). In the northern offshore blocks (A, B, F), redeposition by gravitational mass flow occurred (Rider & Kroon, 2003; Van der Molen, 2004). The Houthem Formation repre-

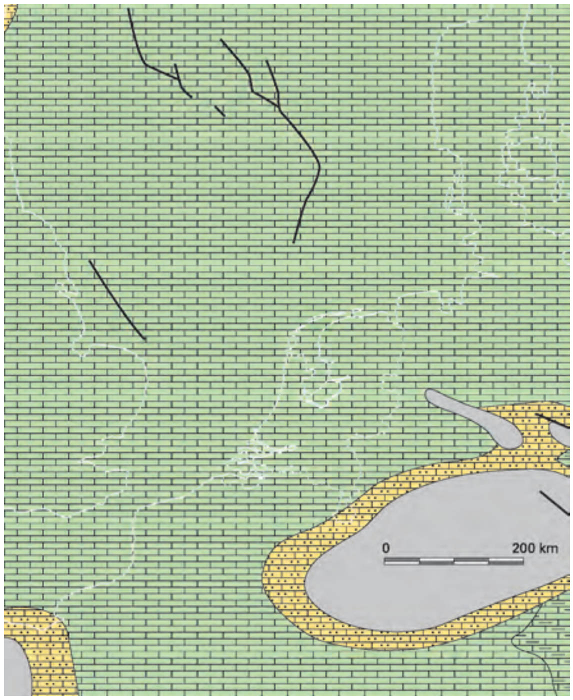

Fig. 17. Paleogeographic map of the Netherlands and adjacent areas during the Senonian-Danian (after Ziegler, 1990). Legend in Fig. 15.

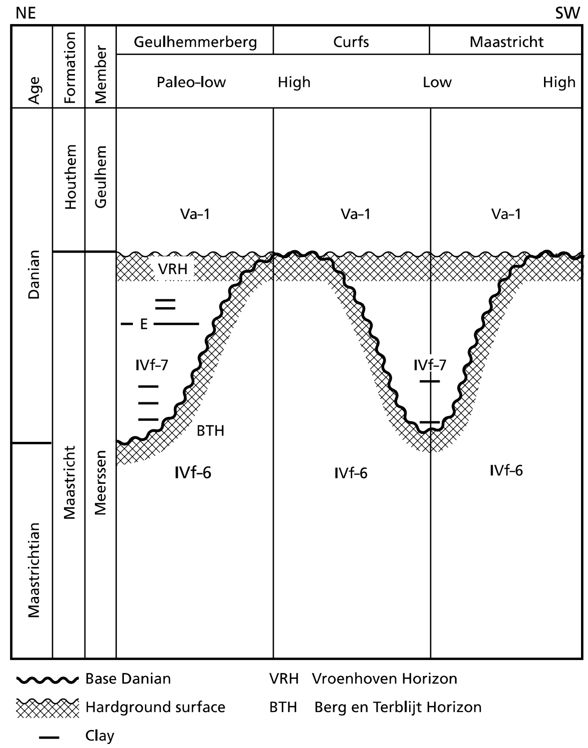

Fig. 18. Stratigraphic relationships in the Cretaceous/Tertiary boundary interval in southern Limburg (after Herngreen et al., 1998).

sents shallow-marine settings near the southern basin edge.

The sedimentation of carbonates ended with an overall regression probably induced by regional lithospheric deformation (Cloetingh, 1986). This was linked to a widespread inversion phase in northwest Europe, known as the Laramide phase, during the Paleocene. In seismic sections, the effects of this phase are often indistinguishable from those of the Subhercynian phase. In the Central Netherlands Basin, the inversion effect removed the Upper Cretaceous completely.

Cretaceous/Tertiary boundary event (KTB)

Smit & Brinkhuis (1996) integrated sedimentological, biostratigraphic, geochemical and paleomagnetic analyses of the marginal-marine Cretaceous/Tertiary boundary section in the Geulhemmerberg galleries in southern Limburg. Herngreen et al. (1998) published additional information from the nearby Curfs quarry. From these, and other papers in Brinkhuis & Smit (1996), the following development is distilled (Fig. 18). The 1.5-m-thick succession of well-bedded IVf-7 calcarenites, deposited above wave base, contains thin 100%-smectite clay intercalations. A remarkable feature is the undulating hardground at its base, with depressions up to 1.5 m deep, in heavily bioturbated IVf-6 calcarenites. This hardground changes laterally, becoming a flat, burrowed surface at the Curfs quarry where it was originally defined as Berg en Terblijt Horizon (Felder & Bosch, 2000). It marks a brief period of non-deposition in the Meerssen Member of the Maastricht Formation. Above this horizon, the initially coarse-grained sediments filling the depressions of the paleorelief represent a thinning-up and fining-up sequence, culminating in the 10 to 15-cm-thick E clay. Bioturbation is conspicuously reduced in the sediments directly overlying the horizon. The inferred process of infilling is episodic, decreasing storm-wave activity in waters deepening to ca. 20 to 40 m, or alternatively a changeover to a more 'shielded' position. The clay layers could have been deposited only during low-energy conditions. According to microfossils, in particular dinoflagellates and benthic foraminifera, and to a lesser degree sporomorphs, the Cretaceous/Tertiary boundary is at the Berg en Terblijt Horizon and not, as traditionally thought, at the overlying Vroenhoven Horizon, separating the Maastricht and Houthem formations. Just below the Berg en Terblijt Horizon, very low, yet relatively slightly elevated Ir-contents are observed. According to Smit & Brinkhuis (1996) the time span between the IVf-6 and IVf-7 sediments, separated by the Berg en Terblijt Horizon, involves deposition of latest Maastrichtian calcarenites, followed by K/T storm or hypercane-related 'washing' of the paleoshelf, and erosion of latest Maastrichtian and earliest Danian sediments, inclusive of an Ir-bearing layer.

Economic geology

Cretaceous deposits are important for the economy of the Netherlands since they contain numerous oil and gas fields (De Jager & Geluk, this volume). In addition, Cretaceous sandstones and limestones have been widely used for building purposes; the limestones largely contributed to the local cement industry.

Oil and gas

Various Lower Cretaceous sandstones are excellent reservoirs, and may be oil or gas-bearing. In the West Netherlands and Broad Fourteens basins, these reservoirs produce mainly oil sourced by the Jurassic Posidonia Shale Formation. In the Lower Saxony Basin, the Bentheim Sandstone of the large Schoonebeek field contains oil generated from the Ryazanian Coevorden Formation. The Vlieland Sandstone deposits in the north of the country are predominantly gas-bearing, the gas being derived mainly from Carboniferous coals.

The onshore Harlingen field is the only Dutch gas field that has been discovered in the Chalk Group (Van den Bosch, 1983; Herngreen et al., 1996). The Hanze field in the F2 block, discovered in 1996, is the first (marginal) Chalk oil-field development in the Dutch North Sea sector (Hofmann, 2002). The Ommelanden and Ekofisk formations constitute the reservoirs.

Water supply

The Waterleiding Maatschappij Overijssel (WMO), including the former Enschede waterworks, derives part of its groundwater from the Bentheim and Gildehaus sandstones at depths down to ca. 150 m. The permeability of these sandstones varies strongly, depending on sedimentological properties, fissures and leaching near the fault planes northeast of the Gronau overthrust. Since 1932, the pumping station Losser has produced ca. 2.4 million m^3/a (VEWIN, 1986; Vermeulen & Stuurman, 1996). More recently, 1.5 million m^3/a have been won from the Bentheim, and ca. 0.4 million m^3/a from the Gildehaus Sandstone, up to January 1, 2001, when the latter exploitation was stopped because of deterioration of quality and environmental reasons (R.A. Kloosterman, WMO, personal communication).

Winning of groundwater in southern Limburg is for a major part from the chalk, which has a high secondary permeability due to fissures and small-scale faults (De Wit, 1988; De Vries, this volume). One of the better producing groups of wells yields some 6 million m^3/a (Provincie Limburg, 1989). The total production from chalk aquifers is ca. 30 million m^3/a (Van Rooijen, 1989). Locally, exploitation from sandstone beds in the Vaals Formation is by means of sumps (NITG, 1999).

Building stones

Since medieval times the Bentheim Sandstone has been exploited in quarries in the vicinity of Bentheim, just across the Dutch-German border. This excellent building stone was used in numerous churches and town halls in northern Germany and the Netherlands (see e.g. Nijland et al., 2003). Other applications include grave and millstones, drinking troughs for cattle, etc. At present only one quarry is in exploitation.

The Gildehaus Sandstone at the Staring quarry in Losser, east Overijssel, has never been exploited; at present it is a geological nature reserve, showing the Losser sandstone (local name), in honour of W.C.H. Staring, the founding father of geology in the Netherlands (Anderson et al., 1975; Anderson, 1985).

Since Roman times, chalk of the Maastricht Formation has been quarried from extensive subterranean galleries in southern Limburg. Several members yield excellent building stones. Since 1965, minor amounts, less than 1500 m^3/a, of limestone have been extracted from these underground networks. The building stone, now mainly used for restoration, comes from the underground quarry Sibbe. In addition, ca. 1.5 Mt/a of chalk is produced in quarries in southern Limburg. Most of it is used in the cement industry where a relatively low carbonate percentage (67%) is sufficient and up to 22% SiO_2 is needed. Other applications are in lime-fertilizer, ceramics, paper ware etc. and require at least 92 to 94% $CaCO_3$ (Felder & Bosch, 2000).

Flint

Along the eastern slope of the Meuse valley near the village of Ryckholt in southern Limburg, flint was mined to produce axes, scrapers, spearheads, etc., during Neolithic times (5000-2000 BC). These mines were discovered in 1891 by the Belgian archeologist Marcel de Puydt. The first mine shaft was cleared in 1914. Silex bed 10 of the Lanaye Member (Gulpen Fm), with a mean thickness of 11 cm, was exploited in a ca. 8-ha-large field with approximately 2000 mine shafts down to 12 m (Felder et al., 1998; Felder & Bosch, 2000).

Flint was used to make fire, and from Roman times onwards it has been extracted as a by-product of open-cast mining for building stone. More recent applications include the use as refractory stone, crusher lining, flint in lighters, etc. (Engelen, 1989).

Clay and sand

The Vlieland Claystone was of economic value in brick and tile-works, particularly in adjacent Germany. During the last decades, however, these quarries have been abandoned. In Aachen and adjacent Belgium, the Hergenrath Clay was exploited, for example for the 15th to 19th century stoneware of Raeren (Belgium); at present it no longer has any economic significance. Prior to 1960 the Aken Sand was dug for use as filling sand; now only a single quarry near Kelmis (Belgium) is in operation.

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Tertiary

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ABSTRACT

During the Tertiary the Netherlands was situated near the southern edge of the North Sea Basin, a large epicontinental sag basin of north-south orientation, situated above older, Mesozoic rifting structures. The Tertiary succession consists mostly of siliciclastic sediments. Its lowermost calcareous deposits, of Danian age, belong to the largely Late Cretaceous Chalk Group. The siliciclastic succession has been grouped together with the Quaternary in the North Sea Supergroup, consisting of the Lower, Middle and Upper North Sea groups. The bases of each of these groups are marked by distinct unconformities. The lower group comprises Paleocene and Eocene, predominantly marine deposits, the middle group includes mainly Oligocene, marine strata, and the upper group consists of the marine to continental, Miocene and younger sediments. The thickness of the supergroup locally reaches up to 2500 m. The highly variable thicknesses of its lithological units reflect the deposition and erosion in different areas during changes in sea-level and extensional as well as compressional tectonic movements. The Tertiary occurs locally at the surface along the country's eastern and southern borders. Economically it is of interest because of its surface mineral resources, its aquifers and perhaps also its so far not exploited gas accumulations.

Keywords: stratigraphy, Paleogene, Neogene, Netherlands, North Sea Basin, basin development

Introduction

Historical review

In general, the Tertiary sedimentary section of the Netherlands has not been an exploration target for oil companies. Moreover, its locally relatively deep position down to 2500 m below the surface makes it an expensive research objective for non-commercial institutions outside the few areas near the country's eastern and southern borders where it occurs at or near the surface. This explains why relatively little systematic research has been conducted. The regional mapping of the Tertiary for hydrogeological and geothermal research by the Geological Survey of the Netherlands in the early eighties forms an exception. Although, its results were never published at the time, they formed the basis for several recent publications on the geology of Tertiary aquifers (Dufour, 2000; Verweij, 2003; De Vries, this volume).

Numerous publications deal with the stratigraphy of the Upper Tertiary section (e.g. Van den Bosch et al., 1975; Doppert, 1980) or with specific topics like micropaleontology, nannoplankton, heavy minerals and geochemistry. However, studies presenting an integrated overview of the entire Tertiary succession in English are scarce and dated (e.g. Keizer & Letsch, 1963; Van Staalduinen et al., 1979; Letsch & Sissingh, 1983; Zagwijn, 1989). Van Adrichem Boogaert & Kouwe (1997) revised the stratigraphic nomenclature of the Tertiary established by the Nederlandse Aardolie Maatschappij and the Geological Survey of the Netherlands (NAM & RGD, 1980), but also concluded that they had to leave many questions unanswered. At present a Working Group on the Cenozoic is amending this nomenclature (Weerts et al., 2003).

In the new lithostratigraphic subdivision, most of the Paleogene formations and members are unchanged but several are renamed, and some units previously defined as members are given formation status. Because the results of the working group have not been formally published, this chapter applies the lithostratigraphy of Van Adrichem Boogaert & Kouwe (1997), which is currently the standard. In Table 1 the Paleogene lithostratigraphic units mentioned in this chapter are compared with those proposed by Weerts et al. (2003). Overviews of the new stratigraphic framework for the Upper Cenozoic have been published in Dutch by De Mulder et al. (2003), and for the Up-

Table 1. The new lithostratigraphic names for formations and members of the Lower and Middle North Sea groups proposed by Weerts et al. (2003) compared with the nomenclature of Van Adrichem Boogaert & Kouwe (1997).

Van Adrichem Boogaert & Kouwe (1997)	Weerts et al. (2003)
Veldhoven Clay Mbr	Wintelre Clay Mbr
Rupel Fm	Rupel Subgroup
Rupel Clay Mbr	Boom Fm
Vessem Mbr	Zelzate Fm and Bilzen Fm
Steensel Mbr	Eigenbilzen Fm
Tongeren Fm	Zelzate Fm
Brussels Sand Mbr	Brussel Mbr
Brussels Marl Mbr	*not described*
Basal Dongen Sand Mbr	Oosteind Mbr
Basal Dongen Tuffite Mbr	Layer within Oosteind Mbr
Landen Clay Mbr	Liessel Mbr
Gelinden Marl Mbr	Gelinden Mbr
Heers Mbr	Orp Mbr

per Pliocene to Quaternary in English by Rijsdijk et al. (2005). The Netherlands Institute of Applied Geoscience TNO included a chapter on the Tertiary paleogeographic and structural development in their atlas dealing with the onshore subsurface geology (NITG, 2004). Based on TNO data, Duin et al. (2006) recently published a series of thickness and depth maps, some of which have been used in this chapter.

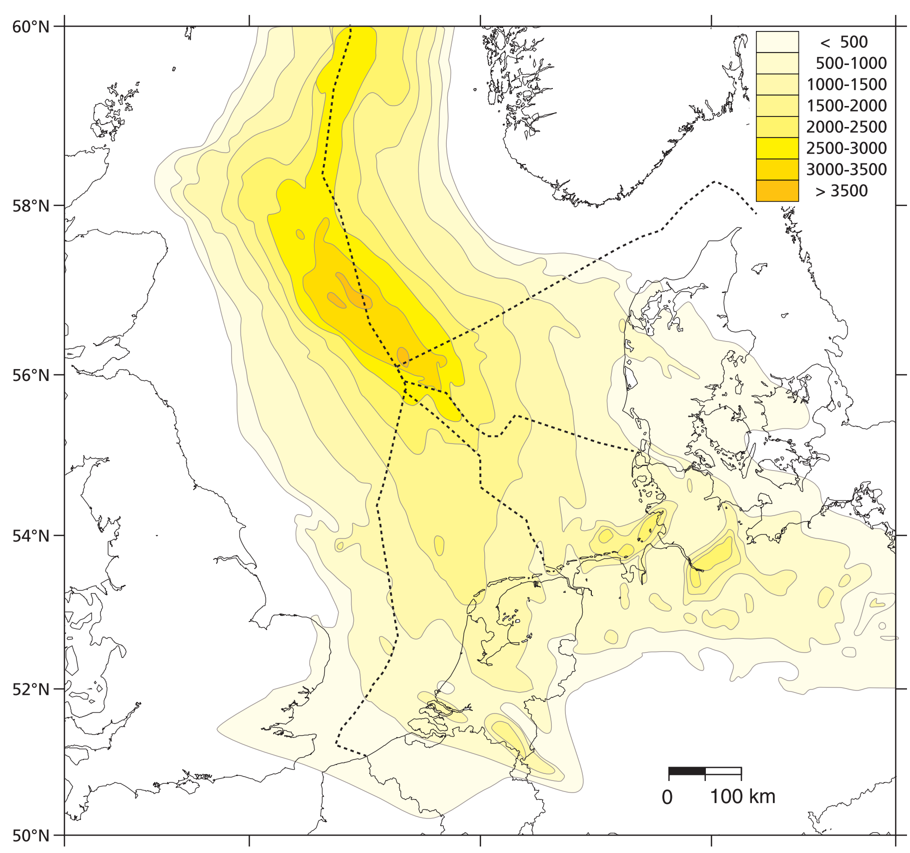
The recurring problem of revising the Neogene stratigraphic scheme is that it has been based on onshore data and that correlation with the offshore region is often enigmatic. However, the increasing availability of high-quality seismic data in the North Sea region enabled various studies on the geometry and stratigraphy of Cenozoic sediments (e.g. Michelsen et al., 1995; Sha et al., 1996; Liu & Galloway, 1997; Clausen et al., 1999; Overeem et al., 2001; Kuhlmann, 2004; Jansen et al., 2004). The overview by Rijsdijk et al. (2005) correlates the younger Cenozoic sediments in the onshore and offshore Netherlands with each other.

Recent studies on the Tertiary of the Netherlands deal with specific areas such as the Broad Fourteens Basin (Wong et al., 2001), the Roer Valley Graben (Verbeek et al., 2002; Michon et al., 2003; Schäfer et al., 2004, 2005) and the southern Dutch North Sea (Overeem et al., 2001; De Lugt et al., 2003). For the in several respects classical Tertiary in Belgium, just south of the Netherlands, Vandenberghe et al. (1998, 2004) presented detailed sequence-stratigraphic overviews. Moreover, there also have been studies modelling the Cenozoic intraplate stress regime (e.g. Kooi et al., 1991; Van Wees & Cloetingh, 1996).

Following Van Adrichem Boogaert & Kouwe (1997), the calcareous, lowermost Tertiary sediments are treated as part of the largely Late Cretaceous Chalk Group (Herngreen & Wong, this volume). At present (e.g. Gradstein et al., 2004), there are discussions about the status of the Quaternary, and consequently the nature and position of a Neogene–Quaternary and Pliocene–Pleistocene boundary. The latter boundary, presently defined in southern Italy at the top of the paleomagnetic Olduvai normal event at 1.77 Ma, was internationally accepted (Aguirre & Pasini, 1985). However, strong biostratigraphic, magnetostratigraphic and climatostratigraphic arguments are against this position as the base of the Quaternary (Zagwijn, 1992; Suc et al., 1997; Partridge, 1997).

The INQUA Commission on Stratigraphy proposes to place the lower boundary of the Quaternary near the Gauss-Matuyama paleomagnetic reversal (base Gelasian)

Fig. 1. Map showing outline of North Sea Basin with thickness (metres) of Cenozoic sediments (excluding Danian); after Ziegler (1990) and Huuse (2002).



at approximately 2.6 Ma (Partridge, 1997). This is near the boundary at base Pretiglian advocated by Zagwijn (1960, 1963, 1974, 1998). Since this boundary has been widely used in the Netherlands we will for the time being continue to apply this definition, which is also used in the chapter on the Quaternary of this book (De Gans, this volume).

Geological setting

The Cenozoic North Sea Basin developed in response to the gradual lithospheric cooling of an underlying Mesozoic rift dome (Ziegler, 1990). It is a large, epicontinental sag basin with a north-south axis, situated above the older, Mesozoic rifting structures. The basin is bordered by the structural highs of the Fennoscandian Shield to the north-east, western and central Europe to the south and the British Isles to the west (Fig. 1). At the end of the Danian the deposition of chalk, which began in the Late Cretaceous, terminated as a result of thermal uplift of the British Isles and compression of the western and central-European Alpine foreland (Ziegler, 1990). These crustal movements are attributed to the Paleocene Laramide phase of the Alpine orogeny. The compression caused uplift in the Netherlands, resulting in severe erosion and the development of a regional unconformity. Following the Laramide phase, siliciclastic sediments of the North Sea Supergroup, consisting of the Lower, Middle and Upper North Sea groups, were deposited in the basin.

Further tectonic events and sea-level fluctuations took place. Figure 2 shows the main structural elements. A major rift system developed during the Eocene and Oligocene in western Europe (Sissingh, 2006). The Rhine Graben system is its north-western offshoot, protruding into the south of the Netherlands (Fig. 3). The Upper Rhine Graben bifurcates at the southern margin of the Rhenish Massif NW-ward into the Lower Rhine Embayment and Roer Valley Graben, and NE-ward into the Hessen depression, forming a triple junction (Ziegler, 1994). The Roer Valley Graben itself is NW-SE oriented, and bounded by the Campine Block in the south-west and the Peel Block and Venlo Block in the north-east (Fig. 4). After uplifting in the Late Eocene, subsidence resumed in the Late Oligocene and continued till the present day (Van Balen et al., 2005).

The Pyrenean unconformity, resulting from tectonic movements and erosion during the Late Eocene and Early Oligocene, marks the boundary between the Lower North Sea Group and the Middle North Sea Group (Letsch & Sissingh, 1983; Van Wijhe, 1987). Later tectonic uplift during Early Miocene times caused most of the southern North Sea area to emerge above sea level once more, due to a regional stress regime induced by the Alpine orogeny. The unconformity between the Middle North Sea Group and the Upper North Sea Group marks this Early Miocene tectonic event which is known as the Savian phase (Letsch & Sissingh, 1983; Wong et al., 2001). Hereafter, regional subsidence resumed again and between the Late Miocene and late Middle Pleistocene, the southern North Sea Basin became the site of one of the world's major ancient delta systems, the Eridanos delta. In the south-east of the basin this delta system built out from the eastern seaboard into the North Sea, fed by west-ward flowing, Baltic and north-German rivers and by precursors of Rhine, Meuse and Scheldt (Bijlsma, 1981; Overeem et al., 2001; Kuhlmann et al., 2004).

In part of the onshore area, the basal Quaternary boundary, at about 2.6 Ma, is marked by a hiatus, which probably resulted from a eustatic drop in sea level, related to the first glacial influence during the Pretiglian which is con-

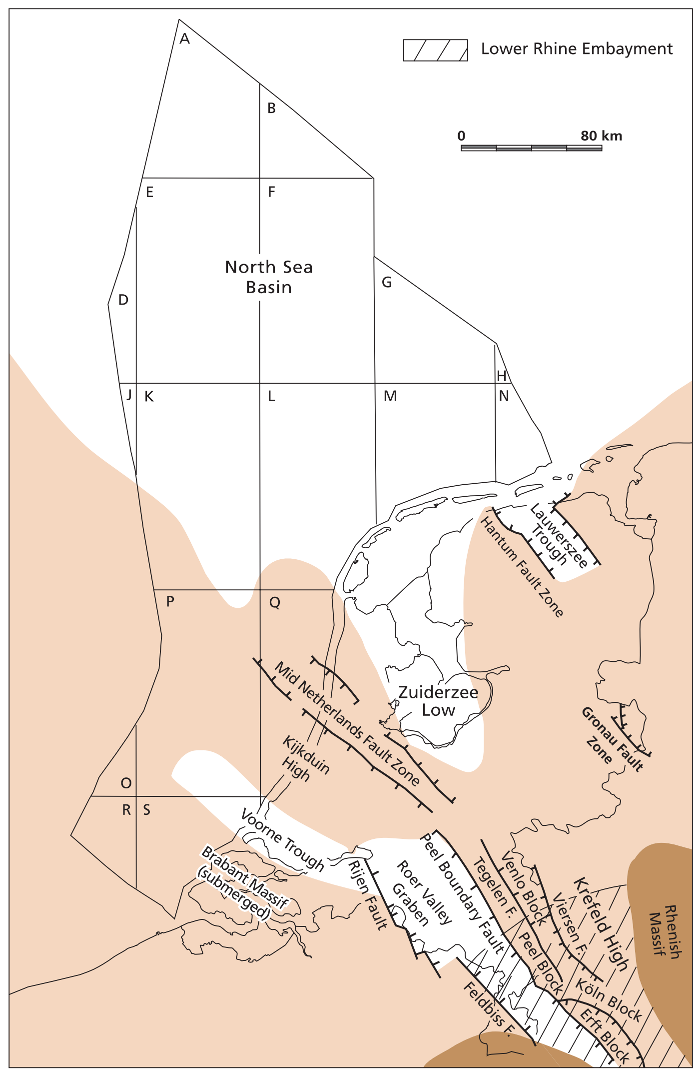


Fig. 2. Map of Cenozoic structural elements in the Netherlands (after Van Adrichem Boogaert & Kouwe, 1993). Dark brown: structural high, subaerial landmass; light brown: platform, intermittently flooded; white: basin.

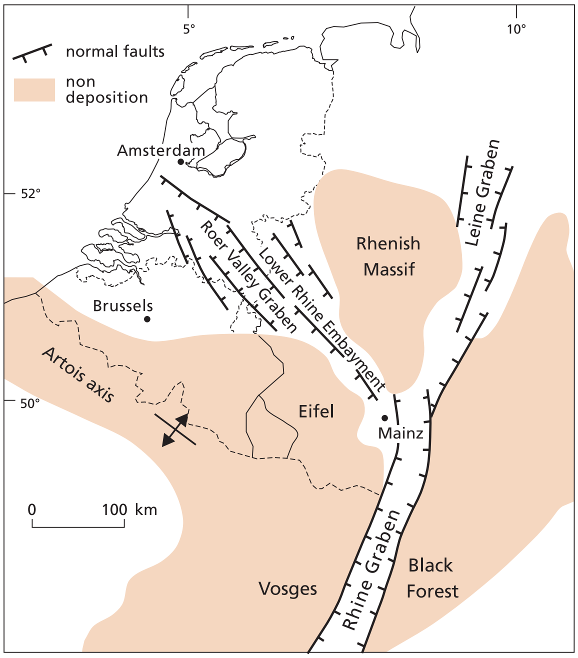


Fig. 3. The Cenozoic graben system in north-western Europe (after Geluk, 1990).

sidered in the Netherlands as the first stage of the Quaternary (Van Adrichem Boogaert & Kouwe, 1997). In the offshore this boundary falls within the prograding delta sequences.

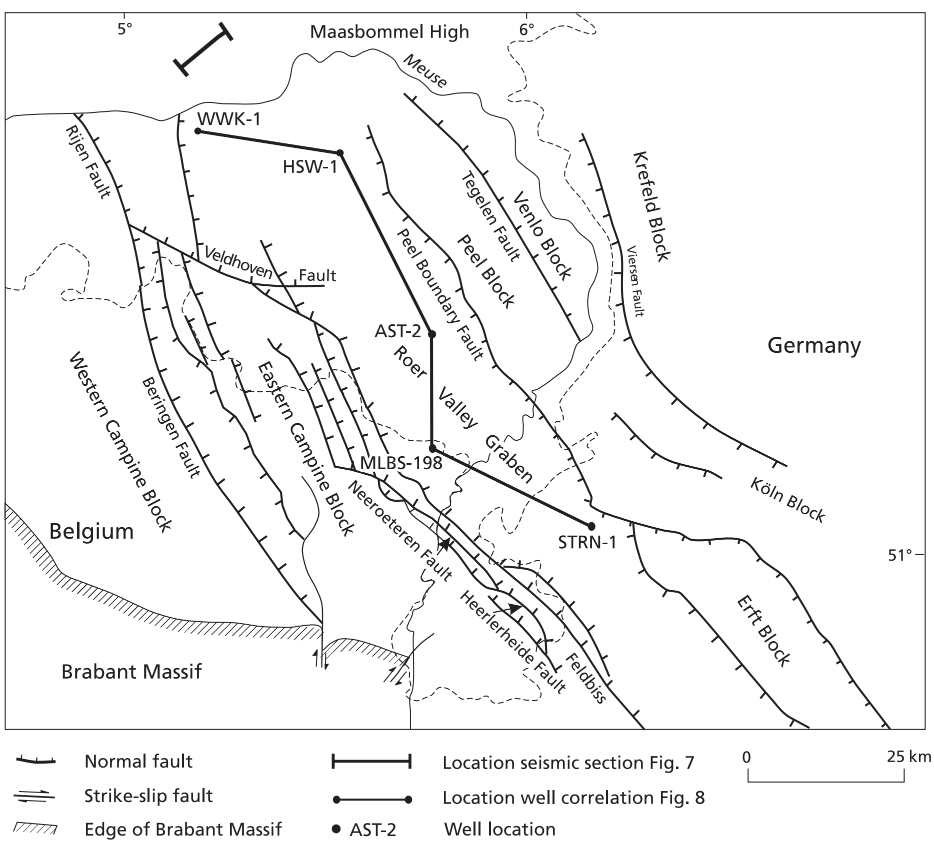
Stratigraphy

The siliciclastic Tertiary succession has been grouped together with the Quaternary sediments in the North Sea Supergroup. Due to its predominantly marine origin, the age determinations and correlations rely in general mainly on marine microfossils (foraminifera and nannoplankton), supported by well-logs, seismic data and sequence-stratigraphy. Continental facies occur only locally near the basin edge.

North Sea Supergroup

The North Sea Supergroup overlies the Chalk Group or older, Mesozoic strata unconformably. Only locally, e.g. in the south of the Netherlands, it rests conformably on the Chalk Group. Figure 5 shows its subcrop in the onshore area. A subcrop map at the base of the Tertiary for the entire Netherlands area was published by Burgers & Mulder

Fig. 4. Main structural elements around the Roer Valley Graben (after Verbeek et al., 2002). Figure also shows locations of seismic profile of Fig. 7 and correlation panel of Fig. 8. Wells: WWK-1 = Waalwijk-1, HSW-1 = Heeswijk-1, AST-2 = Asten-2, MLBS-198 = Molenbeersel-198 and STRN-1 = Straeten-1.



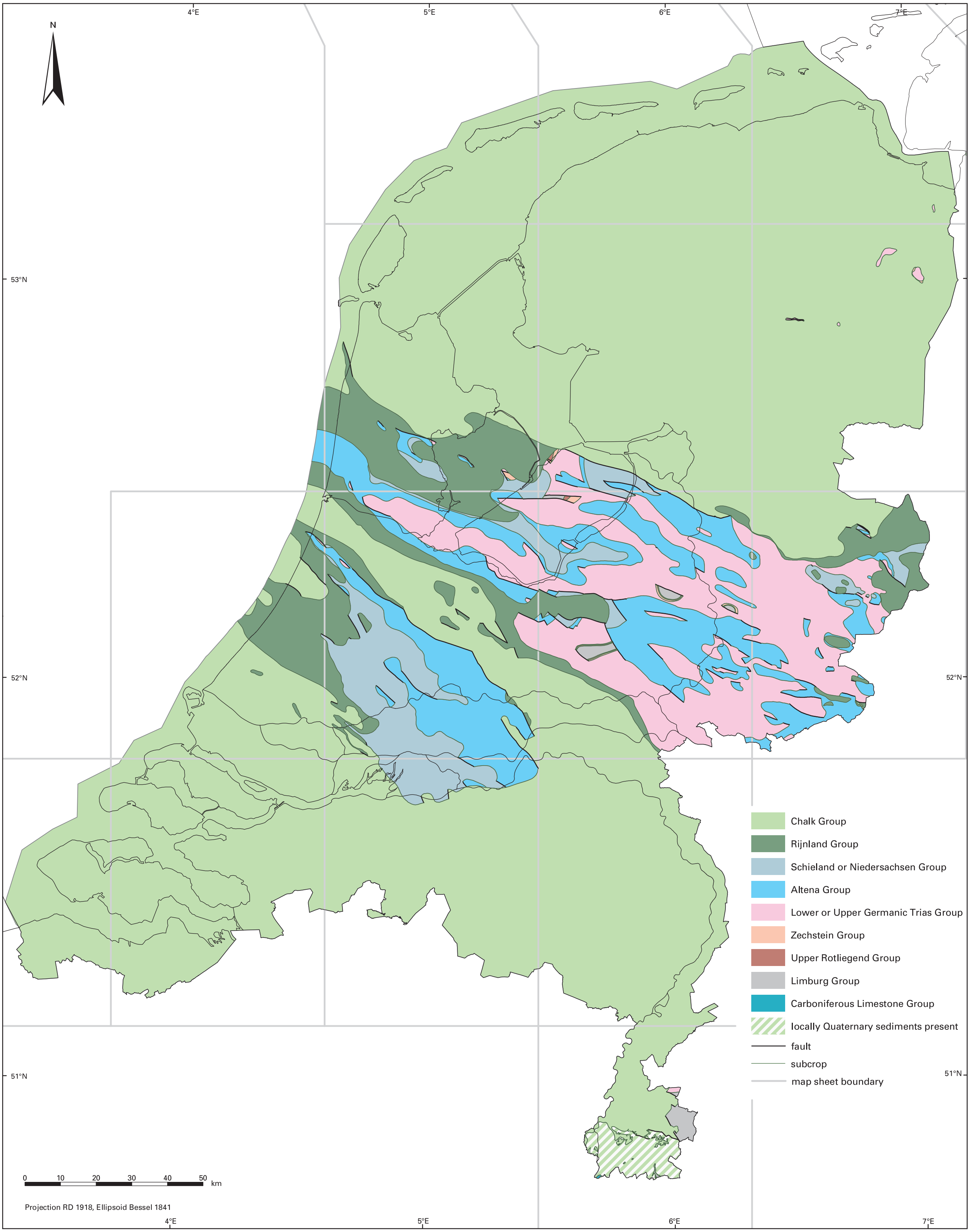


Fig. 5. Subcrop map of the base of the North Sea Supergroup in the onshore area of the Netherlands (after NITG, 2004).

(1991). The depth of the base of the supergroup gradually increases northward to 2500 m in the northernmost part of the Dutch offshore (Fig. 6). The supergroup is divided into the Lower, Middle and Upper North Sea groups. Each of these groups has an unconformable base. The alternation of sands and clays near the southern margin of the North Sea Basin forms the basis of the further subdivision into formations and members. Several of these formations and members are named after localities in Belgium. Van Adrichem Boogaert & Kouwe (1997) show the correlation of the various lithological units into Belgium, Germany and the UK.

The supergroup occurs in virtually the entire country but its thickness is highly variable due to syn-sedimentary block faulting, halokinesis and erosion. A localized maximum thickness of 2000 m is reached in the Roer Valley Graben. Figures 7 and 8 present respectively examples of the seismic expression and well-log characteristics of seismostratigraphic and lithological units in this graben.

The stratigraphy and overall lithological composition of the three groups are reviewed below, following the nomenclature of Van Adrichem Boogaert & Kouwe (1997) for the two lower groups and that of Weerts et al. (2003) for the upper group. Their sedimentary and structural development is dealt with in a next section, which is to a large extent based on Van Adrichem Boogaert & Kouwe (1997). Following these authors, 'Thanetian' is used in this chapter as in Harland et al. (1990).

Lower North Sea Group

The Lower North Sea Group consists mainly of sands, sandstones, marls and clays (Fig. 9). It is subdivided into the Landen and Dongen formations, and is of Thanetian to Bartonian age. It reaches a maximum thickness of approximately 650 m in the northern offshore, thinning distinctly toward the south. An example of the seismostratigraphic subdivision of this group and of its unconformable contact with the overlying Middle North Sea Group is shown in Figure 10. The map in Figure 11 shows the variation in thickness of the Lower and Middle North Sea groups together.

The *Landen Formation*, near the southern fringe of the Early Tertiary North Sea Basin, in the south of the province of Noord-Brabant and in the Netherlands province of Limburg, is characterized by an alternation of fine-grained sands and sandy clays with clays and marls. To the north, this formation consists exclusively of clays with occasional marl intercalations in its lower part. Based on their lithological diversity, five members can be distinguished in the south of the Netherlands. They are named: Swalmen, Heers, Gelinden, Landen Clay and Reussel Member. The Landen Formation attains a maximum thickness of about 100 m in the central-northern offshore area. The formation has a Thanetian age.

The *Dongen Formation* was deposited conformably on the Landen Formation. Like the latter, it comprises sandy deposits along the southern basin edge. These grade northward into marls and partly silty clays. Intercalations of tuff at the base of the formation form in the north a distinct marker horizon which can be recognized on well logs. Six formal members have been distinguished: Basal Dongen Sand, Basal Dongen Tuffite, Ieper, Brussels Sand, Brussels Marl and Asse. The formation reaches its maximum thickness of 600 m in the central-northern offshore, where it is mainly made up by the informal Dongen Clay member. It has an Ypresian to Bartonian age.

Middle North Sea Group

The Priabonian to Aquitanian, i.e. largely Oligocene Middle North Sea Group mainly consists of sands, silts and clays (Fig. 9). The coarser sediments are generally confined to the southern margin of the North Sea Basin. The sediments become finer toward the basin centre. The group rests unconformably on the Lower North Sea Group or older sediments. It comprises the Tongeren, Rupel and Veldhoven formations. It is present throughout the Netherlands except locally in the east and south of the onshore area.

The *Tongeren Formation* consists of a sandy lower part and of a clayey upper part with thin intercalations of sands and lignites. Two members can be distinguished: the Klimmen Member (shallow-marine) and the Goudsberg Member (lagoonal, occasionally coastal-plain). Within the Netherlands, the formation is restricted to Zuid-Limburg where it attains a maximum thickness of 50 m. The formation is of Priabonian and possibly Rupelian age.

The *Rupel Formation* consists mainly of heavy, grey to dark-brown clays with characteristic septaria concretions. The clays grade into sands towards the southern basin margin. In the southern Netherlands the formation has been subdivided into the Vessem, Rupel Clay and Steensel members. For the Twente and Achterhoek areas in the eastern Netherlands, local subdivisions exist (Van Adrichem Bogaert & Kouwe, 1997). The Rupel Formation is present in most of the Netherlands on- and offshore, reaching maximum thicknesses of more than 250 m. It is absent locally along the country's eastern border, on the Kijkduin High and in a few places in the western offshore. Its age is Priabonian to Rupelian.

The *Veldhoven Formation* is developed most typically in the Roer Valley Graben where it comprises a predominantly sandy unit followed by a clay unit that coarsens upward into sands. In this area a distinction can be made into the Voort, Veldhoven Clay and Someren members. The present-day occurrence of the formation is mainly in and around the Roer Valley Graben where it can be as thick as 400 m, and from where it extends northward

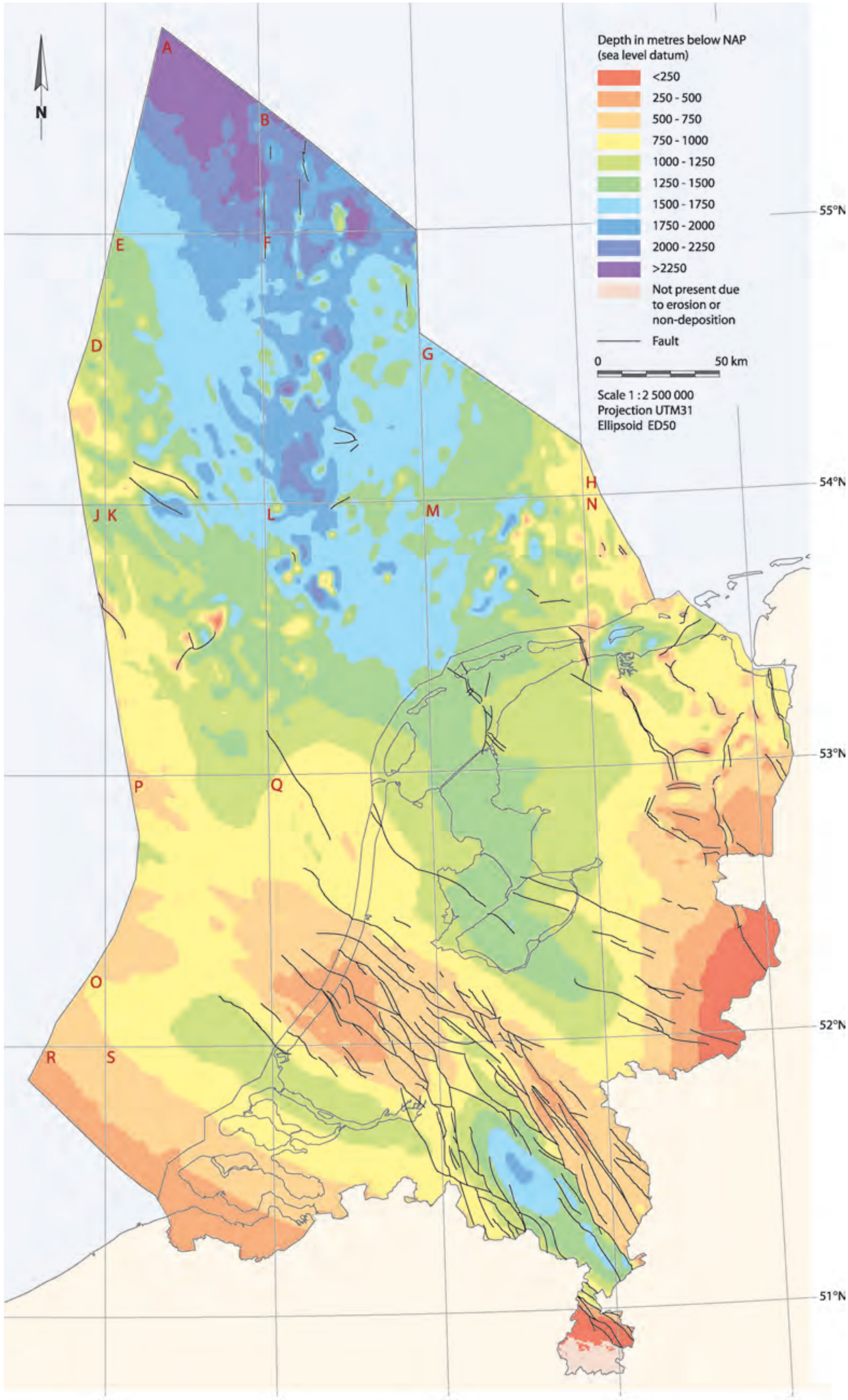


Fig. 6. Depth map of the base of the North Sea Supergroup (Duin et al., 2006).

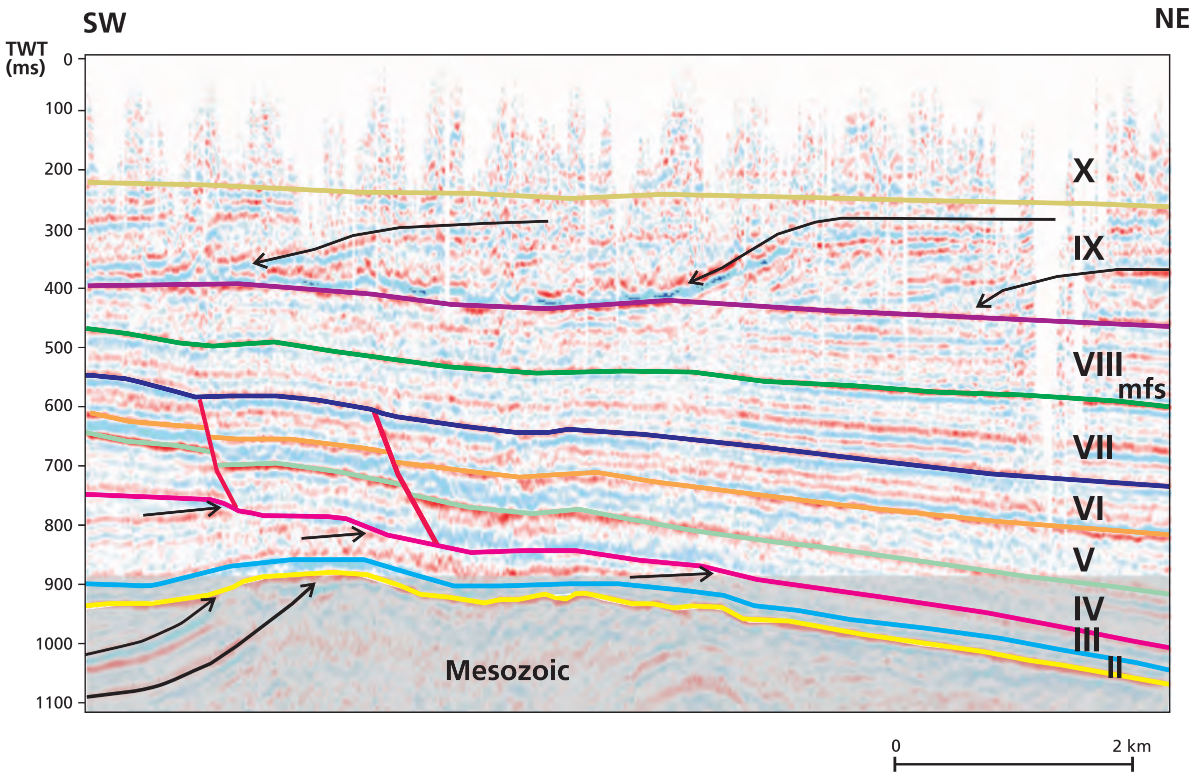


Fig. 7. Seismic section across the Roer Valley Graben, showing Tertiary seismostratigraphic units (after NITG, 2001, and Verbeek et al., 2002). See Fig. 4 for location. Numbered units: II = Landen Fm, III = Dongen Fm, IV = Rupel Fm, V = Veldhoven Fm, VI = lower part of Breda Fm, VII = middle part of Breda Fm, VIII = upper part of Breda Fm, IX = Oosterhout/Kieseloolite Fm and X = various Quaternary formations; mfs = maximum flooding surface. Note the large-scale sigmoid clinoforms in the upper part of the section, characteristic for prograding sedimentation, as is the case with deltaic deposits.

into the south-eastern part of the Zuiderzee Low. Its age is Chattian to possibly Aquitanian.

Upper North Sea Group

The Upper North Sea Group consists of clays, fine to coarse-grained sands and locally gravel and peat or brown-coal seams. The group, of which the base reaches a maximum depth of 1400 m in the Roer Valley Graben and Central Netherlands Basin (Fig. 12), rests unconformably on the Middle North Sea Group or older deposits. It is of Neogene and Quaternary age. Traditionally, the lithostratigraphic subdivision of the Late Tertiary part of the group into the marine Breda and Oosterhout formations and the predominantly continental Ville, Inden, Kieseloolite and Scheemda formations (Fig. 13) was based on information from the onshore area only (Van Adrichem Boogaert & Kouwe, 1997). In the new stratigraphic subdivision of this group, Weerts et al. (2003) distinguish marine, fluvial, glacial and terrestrial sediments, each category consisting of various formations (see also De Mulder et al., 2003; Rijsdijk et al., 2005). In that subdivision both onshore and offshore data have been integrated. The Quaternary formations that belong to the group are dealt with by De Gans (this volume). The group is present over most of the Netherlands. In the extreme east, south-east and south-west, it is locally absent as a result of non-deposition or erosion.

Marine deposits The *Breda Formation* comprises glauconitic sands and sandy to silty clays. Its glauconite content locally reaches more than 50%, which gives this unit its characteristic green to blackish green colour. Based on lithological developments in Zuid-Limburg and the eastern Netherlands, where it occurs locally at or near the surface, the formation has been subdivided into numerous members (Weerts et al., 2003). The formation is present in most of the onshore Netherlands except for some small areas in the extreme east, south-east and south-west (Fig. 14a). It also occurs under the North Sea. The formation attains its maximum thickness of more than 600 m in the Roer Valley Graben. Its age is Miocene, locally also earliest Pliocene.

The *Oosterhout Formation* consists of sands, sandy clays and clays with a glauconite content that is much lower

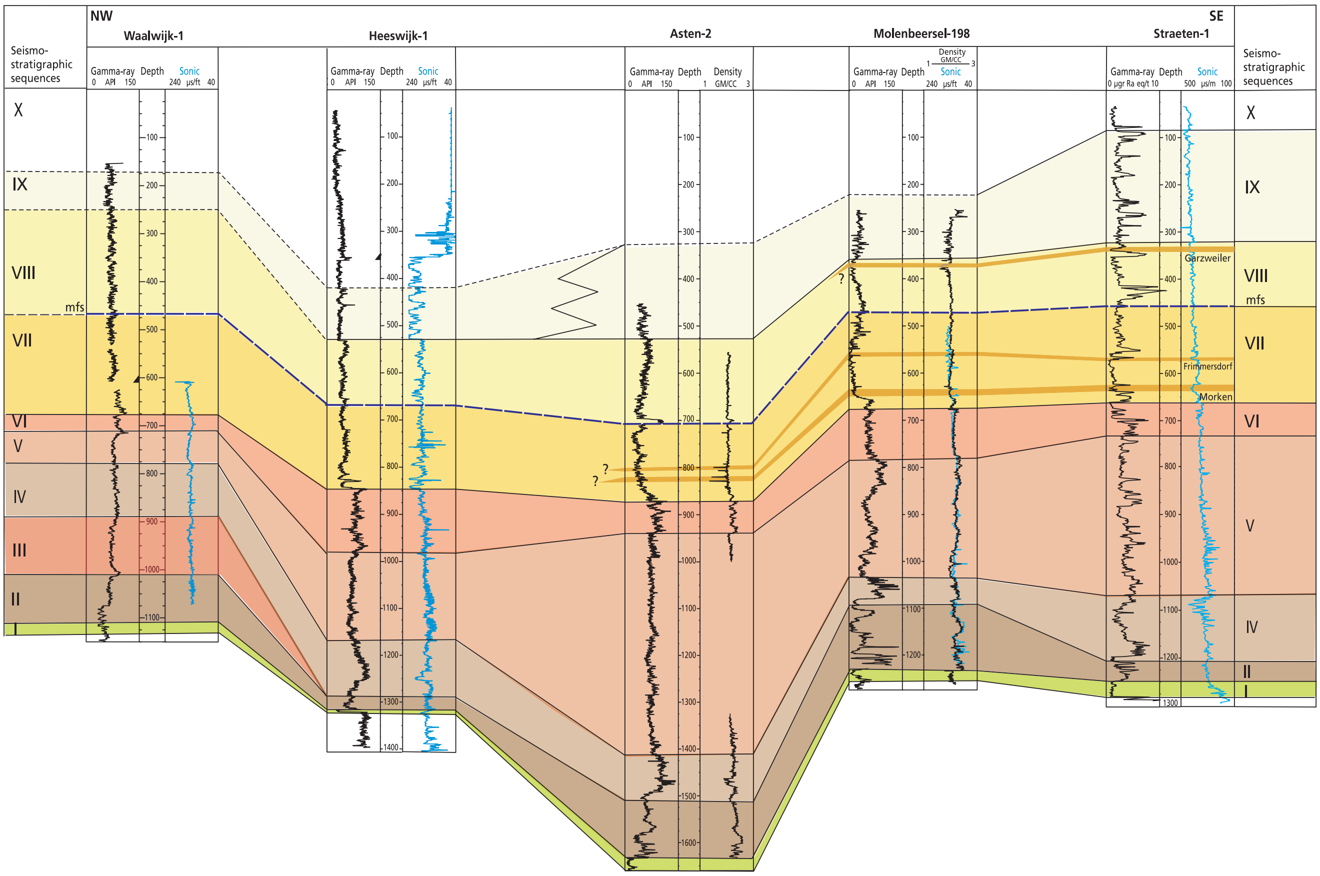


Fig. 8. Log correlation of wells in the Roer Valley Graben (after Verbeek et al., 2002). Depths in metres. See Fig. 4 for location. Numbered units: I = Houthem Fm, II to X and 'mfs' as in Fig. 7. 'Morken', 'Frimmersdorf' and 'Garzweiler' are names of brown-coal seams of the Ville Fm, a lateral equivalent of the Breda Fm (Fig. 13). Between wells Heeswijk-1 and Asten-2, unit IX changes laterally from Oosterhout Fm (NW) into Kieseloolite Fm (SE).

than that of the underlying Breda Formation. In the north-east and south of the Netherlands, the lower part of the formation is characterized by sands that are extremely rich in mollusc shells and bryozoans ('Crag facies', cf. Van Adrichem Boogaert & Kouwe, 1997). In general, the formation gradually changes into the overlying, equally marine, Pleistocene Maassluis Formation. In the eastern Netherlands, the formation intertongues with, and is overlain by continental deposits of the Kieseloolite Formation. The formation is generally present in the country's land area except for some areas in the east, south-east and south-west (Fig. 14b). The formation reaches a maximum thickness of several hundreds of metres. Its age is Pliocene.

The *Brielle Ground Formation* is a seismostratigraphic unit, which forms the lateral continuation of the Oosterhout Formation into the North Sea area. Its generally glauconitic sediments consist of fine sands, clays and sands rich in marine shells. The internal seismic-reflection configuration suggests a delta-front deposit prograding in western and north-western directions (Overeem et al., 2001). Its maximum thickness is 100 m and its age is Pliocene.

Fluvial deposits The *Kieseloolite Formation* consists mainly of light-coloured, fine to very coarse-grained sands, which are locally rich in gravel. Thick, brownish grey and bluish grey silty clays which are generally very carbonaceous, are intercalated. Metre-thick peat or brown-coal layers locally occur within the clay sections. The formation occurs in the south-east of the Netherlands (Fig. 14c) where it attains a maximum thickness of more than 300 m in the Roer Valley Graben. To the west and north it grades laterally into the marine Breda and Oosterhout formations. Its age ranges from Late Miocene to Early Pleistocene (Pretiglian).

The *Peize Formation* consists of medium to very coarse-grained sands which contain gravel locally. Clay beds generally form its basal part. The formation occurs north of the river Rhine (Fig. 14d) where it reaches a maximum

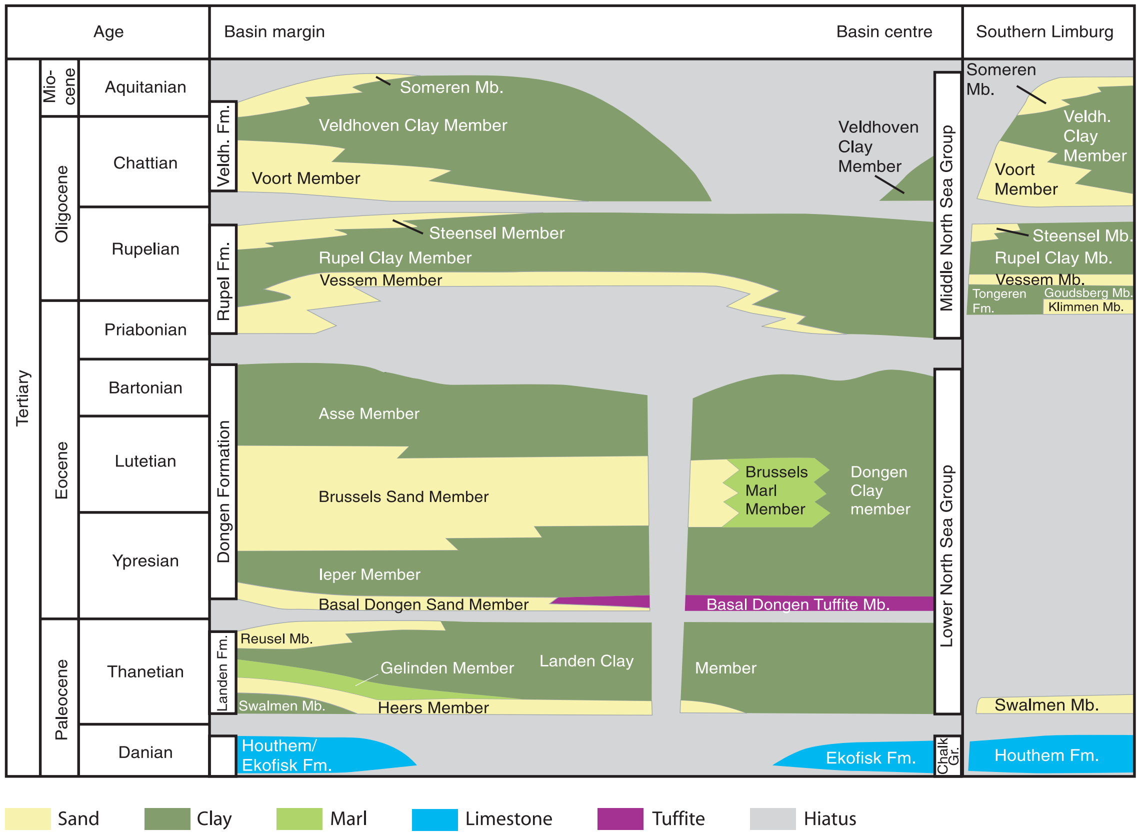


Fig. 9. Stratigraphic scheme of the Lower and Middle North Sea groups in the Netherlands (after Van Adrichem Boogaert & Kouwe, 1997).

thickness of about 150 m. To the west and south it passes laterally into the Maassluis Formation and other deposits. Its age is Pliocene to Early Pleistocene.

Terrestrial deposits The *Ville Formation* comprises brown-coal seams and intercalated sands and gravels. The formation is present mainly in the Roer Valley Graben and in the Erft Fault Block in the Lower Rhine Embayment in Germany (Schäfer et al., 2004, 2005; Mosbrugger et al., 2005). In this embayment the brown-coal layers attain together a thickness of over 100 m (Hager, 1986). Three offshoots of this brown-coal section are present in the south-eastern Netherlands: the Morken, Frimmersdorf and Garzweiler seams (Figs 8, 13). In western direction the formation interdigitates with the marine Breda Formation. Its age is Early and Middle Miocene.

The *Inden Formation* has a geographic distribution similar to that of the Ville Formation. In the type area in Germany it consists of a brown-coal bed of ca. 40 m which laterally interdigates with non-marine and littoral sands, and some clays. This give rise to three coal seams: Friesheim, Kirchberg and Schophoven. The latter extends into the south-eastern Netherlands. In western direction the formation interdigitates with the marine Breda Formation. Its age is Late Miocene.

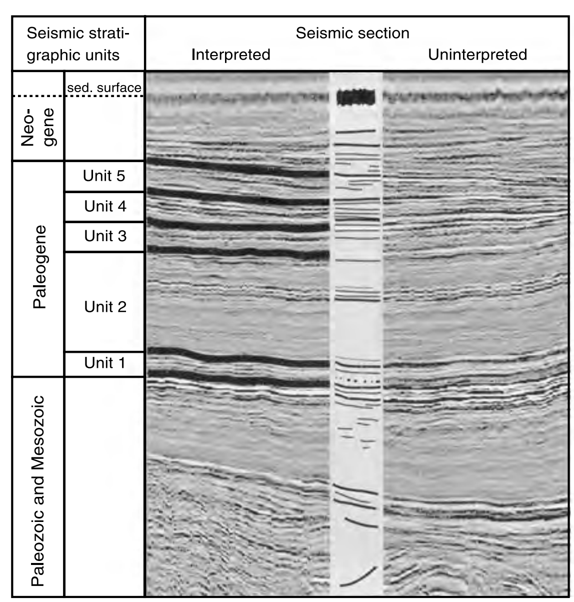


Fig. 10. Section showing seismic stratigraphic units in the Paleogene of the offshore Q block. Unit 1 = Landen Fm, Unit 2-4 = Dongen Fm and Unit 5 = Rupel Fm (after De Lugt et al., 2003).

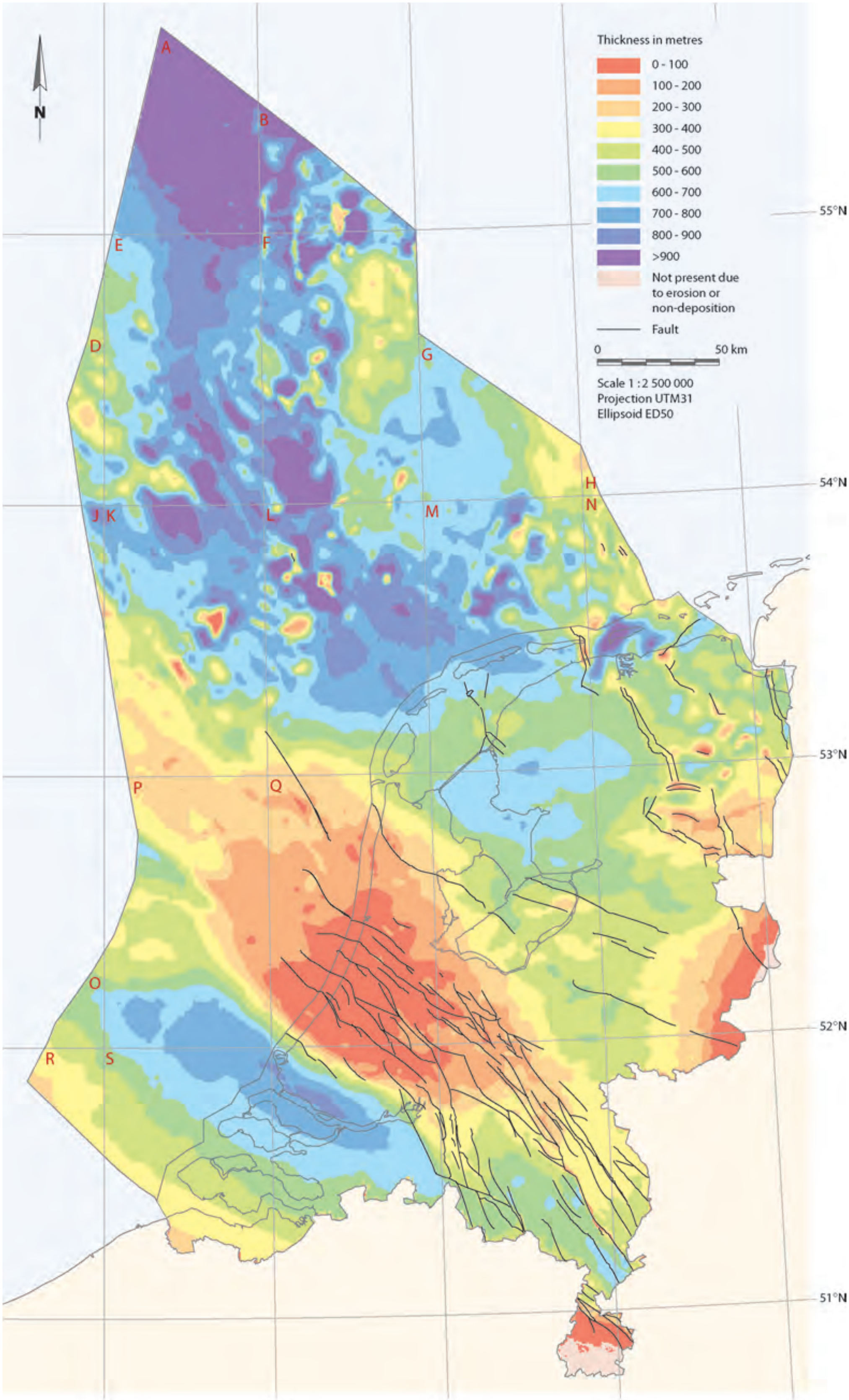


Fig. 11. Isopach map of the Lower and Middle North Sea groups (Duin et al., 2006).

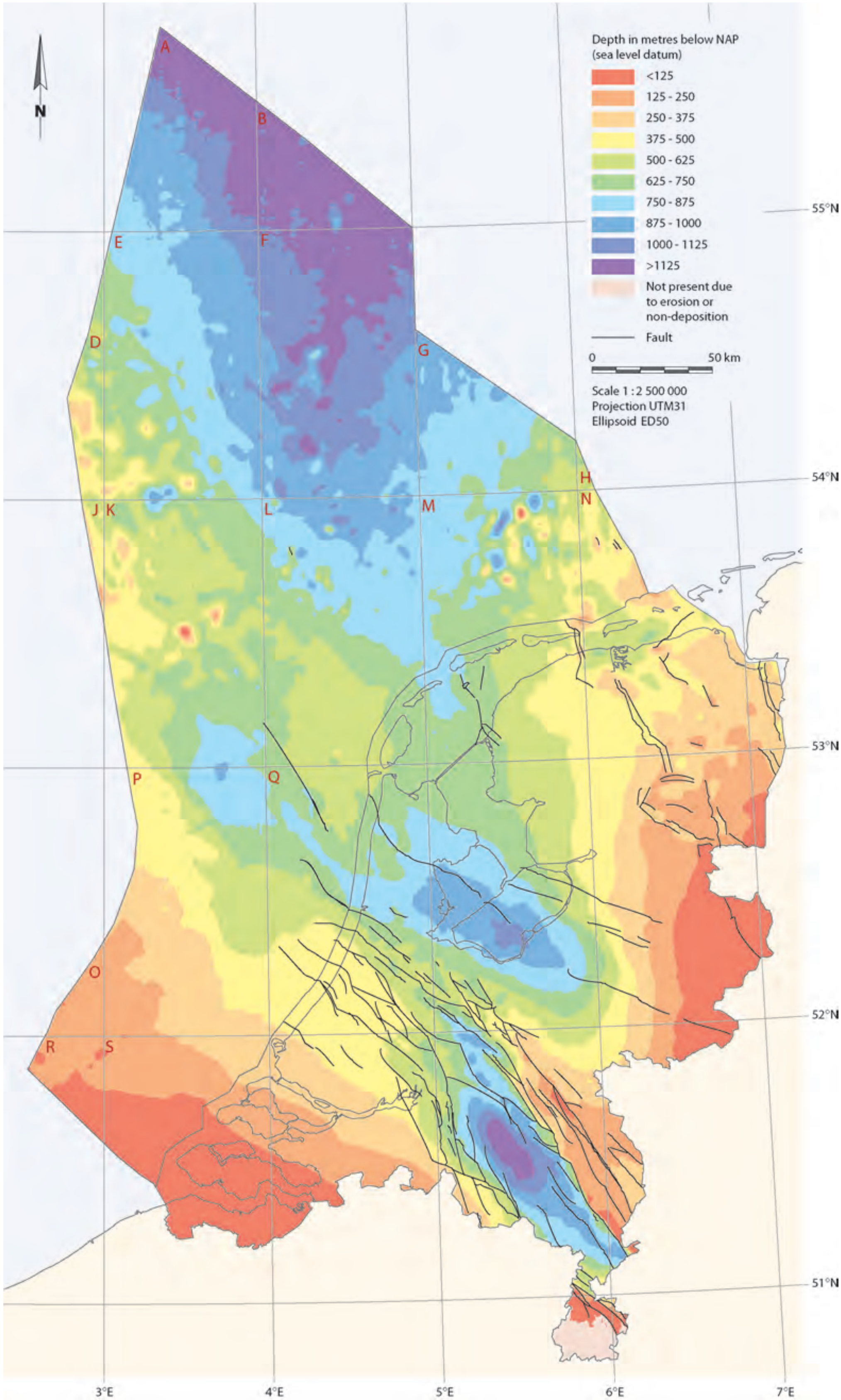


Fig. 12. Depth map of the base of the Upper North Sea Group (Duin et al., 2006).

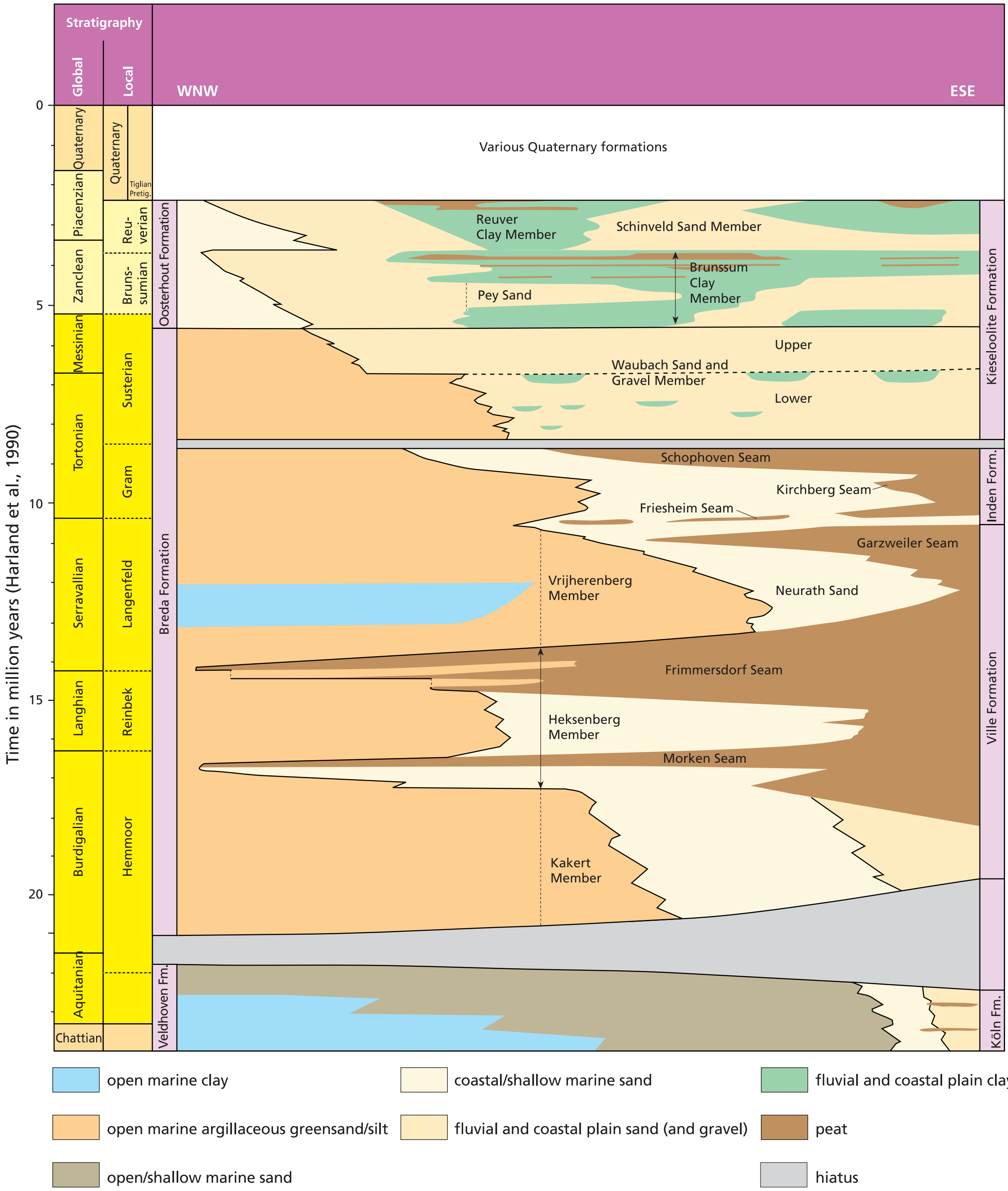


Fig. 13. Stratigraphic scheme of the Upper North Sea Group in the southern Netherlands (after Van Adrichem Boogaert & Kouwe, 1997).

Sedimentary and structural development

Paleogene

During the Subhercynian tectonic phase, which peaked in the Campanian, most of the Netherlands was subjected to local tectonic inversion and subsequent erosion (De Jager, this volume). The uplift was of a gentle nature in both the north and the south of the country (Keizer & Letsch, 1963). A zone in the centre was subjected to more intense deformation and uplift. After peneplanation of the area, a transgression took place and the deposition of the Chalk Group continued during a period of decreased tectonic activity in which the Subhercynian structures were overstepped by Maastrichtian and Danian chalk

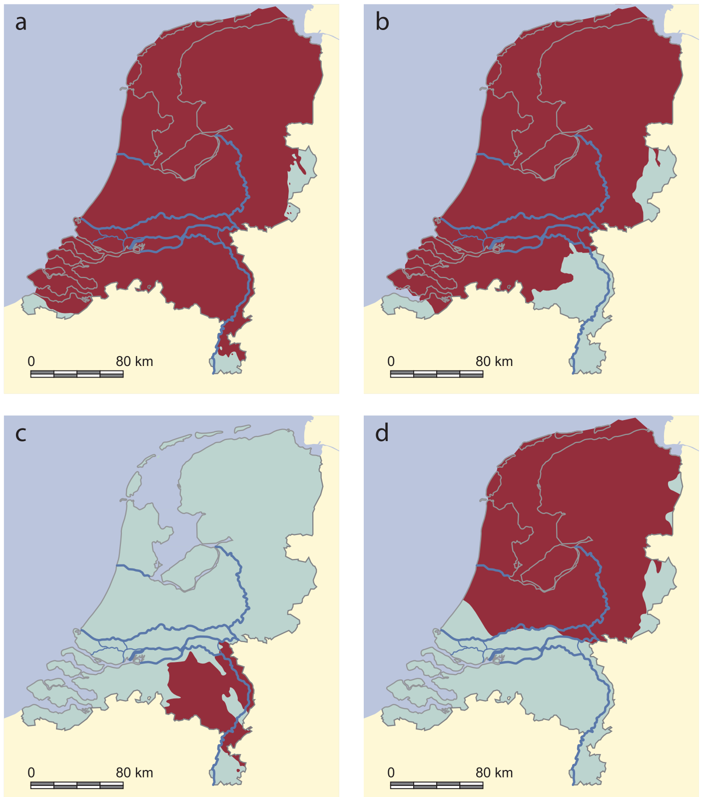


Fig. 14. Maps showing distribution of: Breda Fm (a), Oosterhout Fm (b), Kieseloolite Fm (c) and Peize Fm (d) in the Netherlands onshore (after De Mulder et al., 2003).

(Van Wijhe, 1987; Ziegler, 1990; Herngreen & Wong, this volume). Subsequently, inversion movements accelerated once more during the Paleocene Laramide phase. These movements resulted from reactivation of the fault system that was already active during the Subhercynian inversion, and that was inherited from the Caledonian and Variscan orogenic periods. In combination with a mid-Paleocene sea-level lowstand, these deformations induced major erosion that deeply truncated the uplifted anticlinoria. Much of the Danian chalk was eroded. However, in the adjacent, non-inverted areas, sedimentation was continuous (Van Wijhe, 1987). The Laramide compression of the western and central European Alpine foreland and thermal uplift of the British Isles resulted in a sudden increase in the supply of siliciclastic sediments, which ultimately led to the cessation of chalk deposition (Ziegler, 1990). During a last, minor, rifting pulse in the Late Paleocene, the Voorne Trough, a NW-SE striking basin to the south-west of the Kijkduin High, was formed (Van Adrichem Boogaert & Kouwe, 1993).

The Landen Formation, the first formation deposited after the Laramide phase, is largely transgressive and was deposited during the Late Paleocene (Thanetian). Figure 15 illustrates the paleogeography of the basin during deposition of this formation. It reflects the rapid marine reoccupation of the area from the north (Letsch & Sissingh, 1983). The slightly glauconitic sands of the Heers Member, at some places grading into sandstones, siltstones or claystones, represent transgressive and coastal conditions. The distribution of this member suggests that it filled low-lying areas between local, probably fault-bounded, highs (Wong et al., 2001). According to Letsch & Sissingh (1983), the main sediment supply was from the south-east. Under continuing transgressive conditions the Landen Clay Member was deposited in a shallow-marine environment. The topographic highs that existed during deposition of the Heers Member were now flooded. A short regressive phase ended the deposition of the Landen Formation (Letsch & Sissingh, 1983).

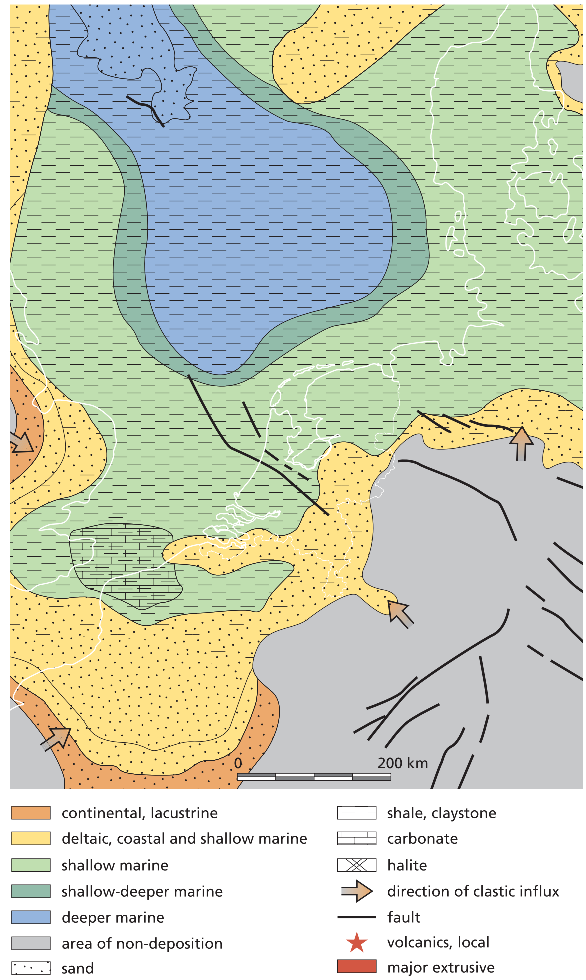


Fig. 15. Paleogeographic map of the Netherlands and adjacent areas during the Late Paleocene (after Ziegler, 1990). Legend shown also applies to Figs 16, 17.

Under renewed transgressive, inner-neritic conditions during the Eocene, the Basal Dongen Sand and Basal Dongen Tuffite members were deposited. The tuffaceous constituents in the latter member are related to the explosive volcanic activity during the transition from Paleocene to Eocene that took place in the northern North Atlantic-Norwegian-Greenland rift zone (Van Bergen & Sissingh, this volume). The generated ash falls form extensive, well-correlatable sections in the North Sea Basin (e.g. Jacqué & Thouvenin, 1975; Knox & Morton, 1988; Nielsen & Heilmann-Clausen, 1988; Ziegler, 1990). The overlying, largely clayey Ieper Member was deposited under outer-neritic conditions. The present-day differences in thickness are most likely due to erosion following uplift during the Pyrenean tectonic phase. De Lugt et al. (2003) suggested that this inversion event, which took place during the Eocene–Oligocene transition, had a much larger intensity than previously assumed. It was accompanied by a regression that resulted in the erosion of the Brussels Sand and Brussels Marl members which had been deposited under shallow-marine conditions in the Lutetian. The present-day distribution of these members reflects erosional remnants only. The overlying clay of the transgressive Asse Member also has been significantly affected by Pyrenean erosion, and is only locally present.

During the Early Oligocene the sea transgressed across the uplifted and eroded areas (Fig. 16) and sedimentation resumed with the deposition of the Tongeren Formation in northern Belgium and Zuid-Limburg. The sands were deposited in a shallow-marine environment, the clays in a lagoonal to coastal-plain environment. Deposition of the Rupel Formation began in the southern regions with the Vessem Member, and elsewhere, in the north, directly with the Rupel Clay Member, which later transgressed southwards over the Vessem Member. The latter member is assumed to be in general a transgressive shoreline deposit (Letsch & Sissingh, 1983; Vandenberghe et al., 2004), with a main direction of supply from the west. The Rupel Clay Member was deposited in a large, relatively shallow sea which extended from Belgium and the Netherlands as far as Denmark and Poland. Towards the end of the Early Oligocene (Rupelian) regression, the deposition of the Rupel Formation terminated.

At the beginning of the Late Oligocene (Chattian) the fault system of the Roer Valley Graben was re-activated (Zagwijn, 1989; Geluk et al., 1994). This caused an accelerated subsidence in the graben and resulted in great differences in the thickness of Upper Oligocene and younger sediments between the graben and adjoining horst blocks. The clays and sands of the Upper Oligocene Veldhoven Formation were deposited in a shallow-marine environment. The formation is mainly preserved in the Roer

Fig. 16. Paleogeographic map of the Netherlands and adjacent areas during the Oligocene (after Ziegler, 1990). For legend see Fig. 15.

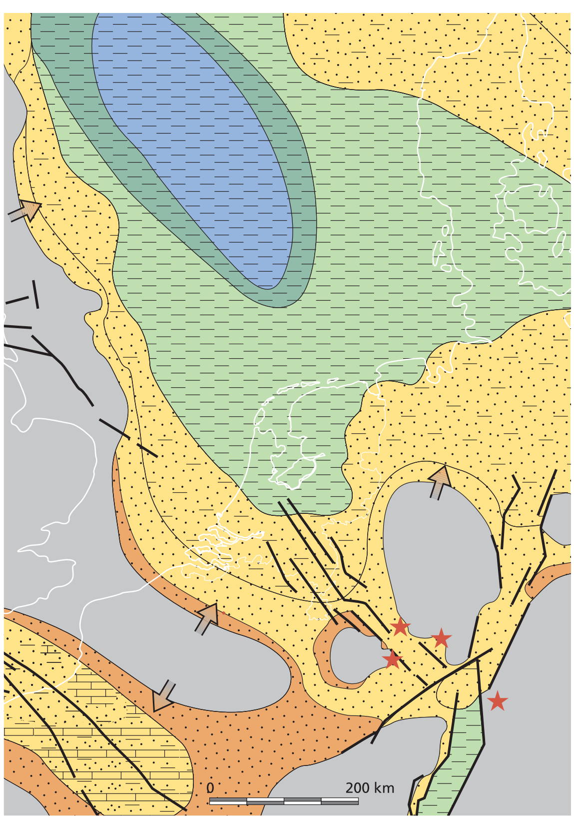
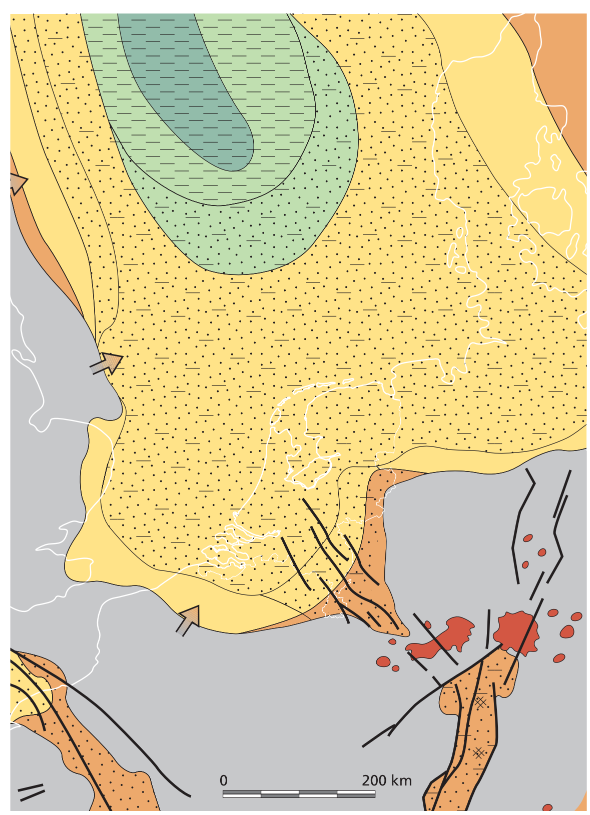


Fig. 17. Paleogeographic map of the Netherlands and adjacent areas during the Mio-Pliocene (after Ziegler, 1990). For legend see Fig. 15.



Valley Graben and adjacent areas but elsewhere it has been eroded during the Savian phase at the end of the Oligocene as a result of regional uplift and a globally low sea level.

Neogene

The Neogene is a period of rapid filling, and consequently shallowing, in the North Sea Basin. The Savian phase, in conjunction with a major sea-level lowstand, culminated in erosion during the Oligocene–Miocene transition. This regressive trend with a number of superimposed transgressions can be recognized in the southern Netherlands (Van Adrichem Boogaert & Kouwe, 1997). The regression led to the westward extension of fluvial sedimentation and peat formation (Ville and Inden formations) into the Roer Valley Graben (Fig. 17). In the deltaic transitional area to the marine environment (Breda Fm) situated to the west, marine marshes with mangroves developed, and widespread peat formation took place. This resulted in thick sections of brown coal, e.g the Morken and Frimmersdorf seams, which are tongues of a massive brown coal deposit (Figs 8, 13; Hager, 1986; Schäfer et al., 2004, 2005; Mosbrugger et al., 2005). During lowstands, the continental deposits built out over the marine depositional area of the Breda Formation (Heksenberg Mbr).

In the Middle Miocene, another delta system started to evolve from the east, supplying sediments in the area of the German Bight and the adjacent part of the North Sea Basin (Bijlsma, 1981; Michelsen et al., 1995; Clausen et al., 1999; Overeem et al., 2001; Kuhlmann et al., 2004). Progradation of the delta resulted in several major sequences within the Upper Cenozoic section in the North Sea Basin (Michelsen et al., 1998). One major sequence boundary is observed within the upper part of the Middle Miocene: the Mid Miocene Unconformity (MMU; Huuse & Clausen, 2001). It is a transgressional surface that is characterized by sediment starvation and/or condensation during the late Middle Miocene (Huuse et al., 2001).

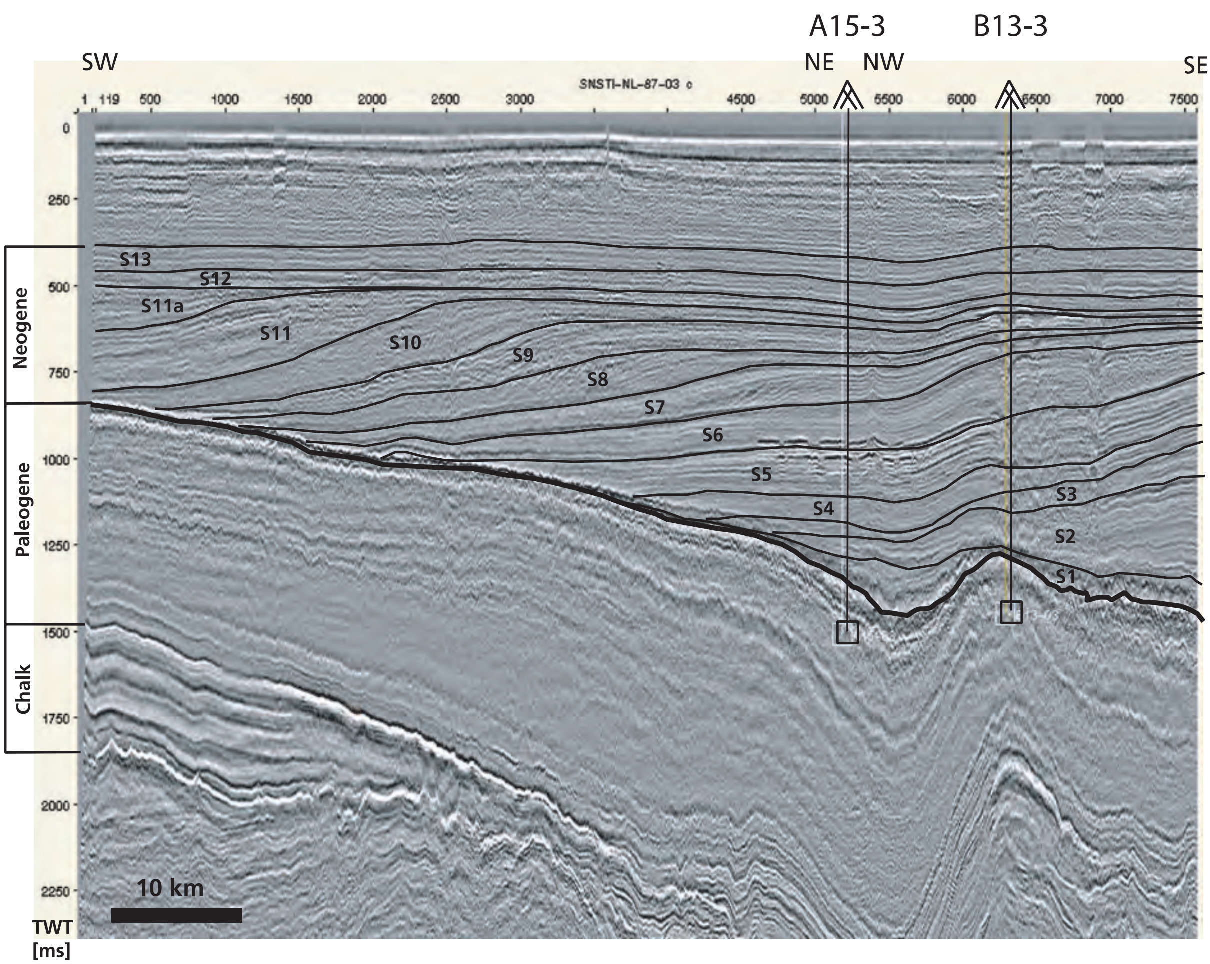


Fig. 18. Seismic section in northern offshore blocks A, B and E, showing prograding character of Upper Cenozoic deposits. Thirteen seismic units (S1 to S13) are indicated. The lower boundary of the interpreted interval is the Mid Miocene Unconformity (MMU); after Kuhlmann (2004). The interval below the MMU is mainly Paleogene but comprises locally, e.g. in well A15-3, a thin Miocene upper part.

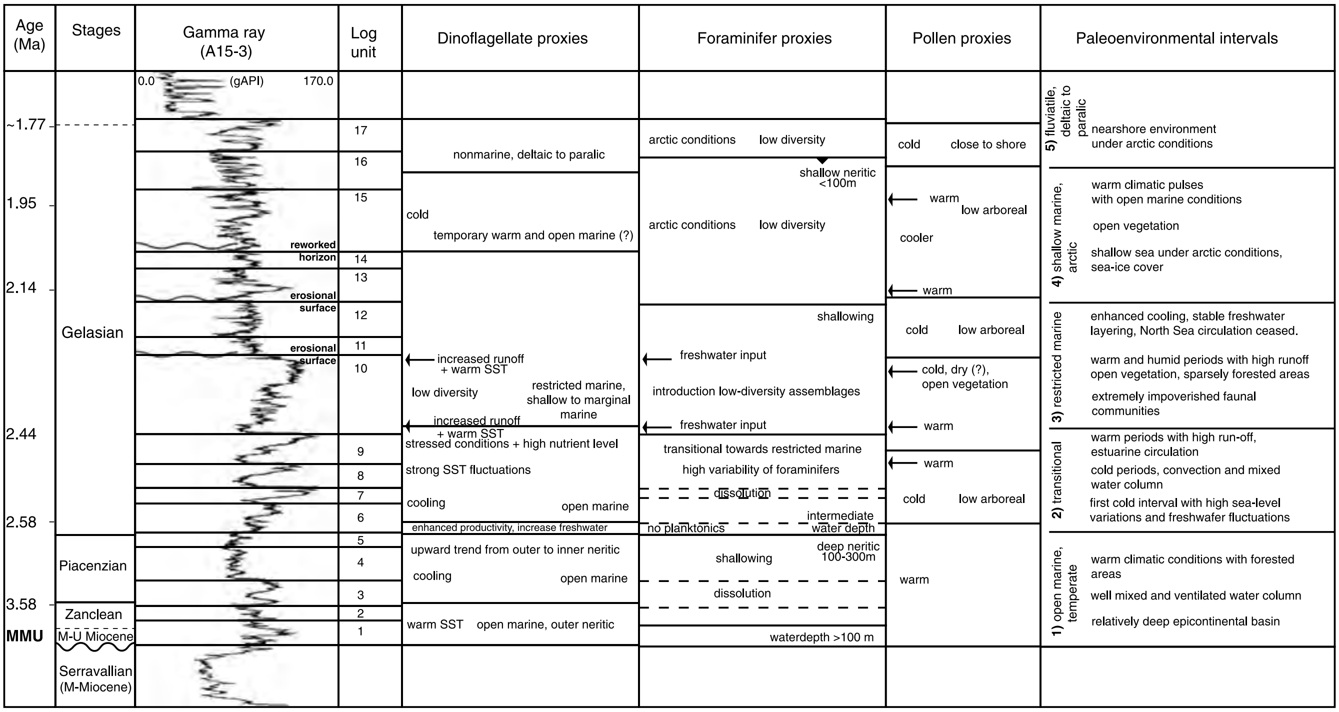


Fig. 19. Proxies of dinoflagellates, foraminifers and pollen for paleoenvironment interpretation compared with gamma-ray response of Late Cenozoic interval of well A15-3 in the northern offshore (Fig. 18). The interval covers the change from relatively warm conditions in the lower part of the 'Pliocene' (Zanclean to Piacenzian) to the onset and intensification of northern glaciation in the upper (Gelasian) part, which in the Netherlands is traditionally considered as lower Pleistocene. High gamma-ray values correspond with relatively cold periods whereas low values represent warmer periods (after Kuhlmann et al., 2006). Thickness of interval shown is ca. 1400 m. SST: Sea Surface Temperature.

The unconformity appears on seismic profiles as an onlap surface towards the north-east of the North Sea Basin (Fig. 18), as a prominent downlap surface within the southern North Sea region, and as a conformable surface in the depocentre, the 'Central Trough' (Cameron et al., 1993; Kockel, 1995; Huuse & Clausen, 2001). For the Dutch North Sea sector, Kuhlmann et al. (2006) dated the sediments below the unconformity to be Middle Miocene or older, on the basis of foraminifer data. Evidence from dinoflagellates indicates a Middle Serravallian age, approximately 12 Ma (Middle Miocene) or older below the unconformity, and a Middle to Late Miocene age directly above it (Fig. 19). The onset of sedimentation onto the unconformity is estimated at 12.4 Ma for the Danish sector (Michelsen et al., 1998) and at 10.7 Ma for the German sector (Streif, 1996). In the southern Norwegian sector the unconformity is dated at 11.2 Ma (Eidvin et al., 1999). Hence, the MMU is younger than the Savian unconformity and therefore it does not separate the Middle North Sea Group from the Upper North Sea Group; it is an unconformity within the latter group.

The Miocene record on the whole is incomplete and locally strongly condensed which makes correlation with the standard sea-level curve of Haq et al. (1988) problematic. During the latest large transgression in the Middle Miocene the sea covered almost the entire Netherlands, except for some areas along the eastern frontier. The glauconitic Breda Formation was deposited under inner to outer-neritic conditions (Fig. 19). According to Letsch & Sissingh (1983) the formation was widely eroded during a tectonic phase at the Miocene–Pliocene transition. Hereafter the final Neogene sedimentary cycle began with deposition of the Oosterhout Formation.

The southern delta in the Pliocene was formed by a precursor of the Rhine which transported eroded material from the uplifted Rhenish Massif as far as the Roer Valley Graben and the Venlo Block (Fig. 2). At the same time, the tectonic activity along the Peel Boundary Fault also augmented and the Roer Valley Graben subsided more and more. The accumulating sand and gravel deposits constitute the Kieseloolite Formation. Locally, this fluvial area was replaced by lakes, where clay and peat accumulated.

A huge delta system existed about 9 Ma ago in the south-east of the North Sea Basin. It was part of an ancient fluvio-deltaic system that prograded through northwest Europe due to simultaneous uplift of the Fennoscandian Shield and accelerated subsidence of the North Sea Basin (Vinken, 1988; Streif, 1996). The dimensions of its drainage area and the thickness of the deltaic deposits are comparable to those of the present Orinoco delta system in Venezuela (Coleman & Roberts, 1989). Its proto-delta in Poland has been named after the 'Eridanos', a mythical amber-bearing river in the north of Europe (Kosmowska-Ceranowicz, 1988), and this name appears appropriate for the entire river and delta system. The system's Upper Miocene and Lower Pliocene sediments consist of thick, laterally confined, sequences which have been deposited in deep-marine water (Figs 18, 19). In the course of time the straight, wave-dominated delta front became gradually lobate and river-dominated. Biostratigraphic data show that the lower, Zanclean to Piacenzian, part of the succession was deposited under relatively warm conditions, while the upper, Gelasian part represents the onset and intensification of northern glaciation (Fig. 19, Kuhlmann et al., 2006). A dramatic sea-level lowering affected the Pleistocene deposition and resulted in strong progradation and submarine mass movements as evidenced by erosive channels and scars and depositional fans in three-dimensional seismic data. From the Pleistocene onwards, more frequent sea-level and climatic changes disrupted the system (Overeem et al., 2001; Kuhlmann, 2004).

Economic geology

Gas

Small amounts of gas have been recorded from Eocene sections above deeper gas-bearing reservoirs in the onshore area of the Netherlands. They have not been exploited so far. These gas-bearing horizons belong to the Basal Dongen Sand and Basal Dongen Tuffite (De Jager & Geluk, this volume). Larger gas accumulations occur in Pliocene-Pleistocene sediments in the A and B blocks in the northern offshore at depths of 400 to 700 m (Schroot & Schüttenhelm, 2003; Kuhlmann, 2004; De Jager & Geluk, this volume). Due to technical complications related to unconsolidated sediments and low reservoir pressures, the production of these reservoirs has not yet been initiated.

Gravel, sand, clay, lignite and phosphorite

Eocene to Pliocene sands and clays have been exploited for many years in the eastern and southern Netherlands. Most of the pits are closed now, but some are still active, for instance those in Zuid-Limburg, in which Miocene silica sands (Heksenberg Mbr) are exploited, and a pit in east Gelderland which exploits Oligocene Rupel Clay (Van der Meulen et al., this volume). Pliocene gravels and Miocene lignites have been extracted in Limburg. In the eastern Netherlands, phosphorite nodules, occurring in the Rupel Formation, have been exploited during the First World War (Pannekoek, 1956). The phosphorite nodules in Pliocene sediments in the south-western province of Zeeland contain insufficient uranium and phosphorus to start exploitation (Harsveldt, 1973).

Aquifers

The sandy sections of the Breda and Oosterhout formations form major aquifers in the south of the Netherlands (De Vries, this volume). To a lesser extent the Vessem and Brussels Sand members also contribute to the production of groundwater.

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Quaternary

W. de Gans

ABSTRACT

Most superficial rocks in the onshore and offshore Netherlands are Quaternary sediments. Onshore they comprise from west to east in essence a Holocene barrier system, interrupted by estuaries and tidal inlets, followed by a Holocene coastal plain, with in the north an intertidal area, and more inland, cut by Holocene rivers, a low and mostly flat area of largely sandy Pleistocene deposits. Due to continuous subsidence in the southern North Sea Basin the Quaternary is locally over 800 m thick. The alternation of glacial and interglacial periods during the Quaternary resulted in strong variations in sea level, shifts of coastlines and hence in a large variety of facies. Glacigenic deposits with a northern provenance, marine and lagoonal deposits with a largely north-western provenance and fluvial deposits with an eastern to southern provenance are stacked upon each other. Mass-flow, eolian and peat layers occur locally. Common reworking of sediments further complicates the stratigraphic framework. In recent times, man has become an important geological agent, causing fundamental changes in depositional environments and landscapes. In the past, Quaternary peat served as an important fuel. Quaternary surface minerals presently exploited are sand, gravel and clay. Aquifers in Quaternary sediments contribute much to the production of drinking water.

Keywords: Netherlands, sea-level variation, glaciation, stratigraphy, paleogeography, human influence

Introduction

The sediments at the surface of the Netherlands, covered as they are by vegetation, manmade constructions or water, are almost exclusively Quaternary (Fig. 1). Older rocks occur only locally at or near the surface along the country's eastern and southern borders. The Quaternary sediments are composed of a Holocene coastal plain in the west and north, and to the east and south of sandy Pleistocene areas cut by Holocene rivers.

Numerous investigations have been devoted to these sediments. This chapter presents an overview of their outcome. It briefly reviews the historical development of the studies and the applied stratigraphy, then the overall geological setting, and subsequently discusses the sedimentary and paleogeographic development of the Pleistocene and Holocene, and the influence of man as a geological agent. It ends with short sections on economic geology and geo-conservation.

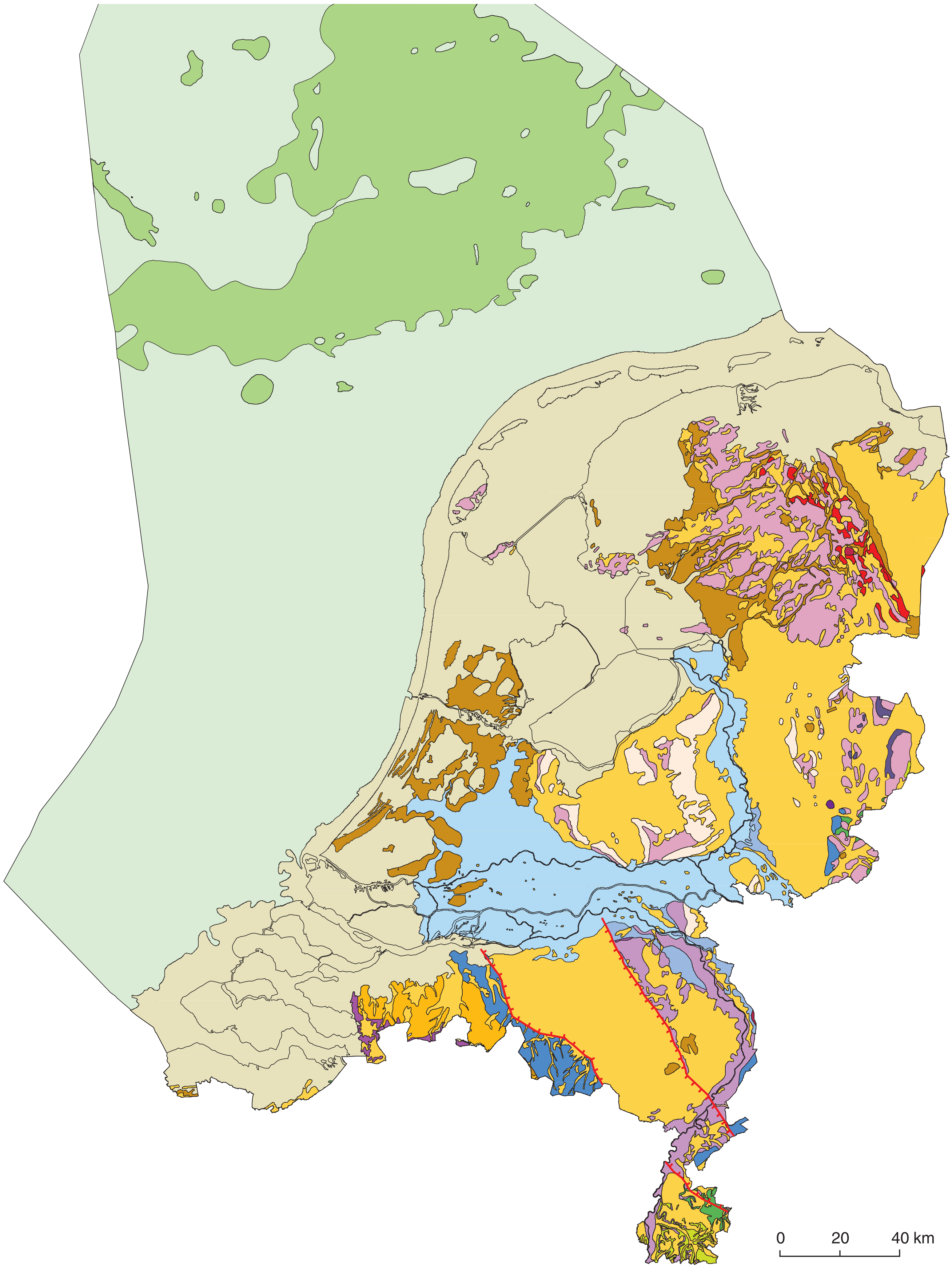
Previous studies

Studies of the Quaternary geology of the Netherlands started in the early 18th century. In those days, inhabitants of the Netherlands became interested in for instance the origin of erratics which were found in large quantities in the north-east of the country. Two main theories were discussed: whether they had been transported by some medium from elsewhere or had grown in the soil.

Interest in the Dutch Quaternary was stimulated by the publication in which Sedgwick in 1825 subdivided the period after the Tertiary in Diluvium (deposits resulting from a major flood) and Alluvium (latest deposits; see Van der Woud, 1998). Soon after, Staring (1833), in his dissertation 'Specimen academicum inaugurale de geologia patriae', invited his readers to study the Diluvial and Alluvial deposits in the Netherlands. In 1844 he published his 'Proef ener geologische kaart van de Nederlanden' (Specimen of a geological map of the Netherlands), later followed by geological maps to the scale 1 : 200 000, which were completed in 1868 (Oele, 2001). In these maps the basic classification of the superficial strata was given, which is still more or less valid. It was Staring also, who introduced the term 'grondsoort' (soil type).

In 1860 Staring published his popular 'Schoolkaart voor de Natuurkunde en Volksvlijt van Nederland' to the scale 1 : 200 000 (School map for the natural science and industry of the Netherlands) which was used in many schools for many years.

In 1837, from studies in the Swiss Alps, the concept 'Ice Age' was postulated as a period in which a major part of Europe had a polar climate and was partly glaciated. In 1847 it became clear that Scandinavian glaciers once covered far larger areas than they do today. This concept was introduced for the Netherlands by Staring (1854) and published in his work 'Het Diluvium van Nederland' (The Diluvium of the Netherlands). As a result the 'Nederlandsche Staatscommissie voor de geologische beschrijving en kartering van Nederland' (Netherlands State Commission for the geological description and mapping of the Netherlands) could in 1854 subdivide the Diluvial deposits in the Netherlands into Meuse deposits, Rhine deposits and deposits formed by Scandinavian glaciers.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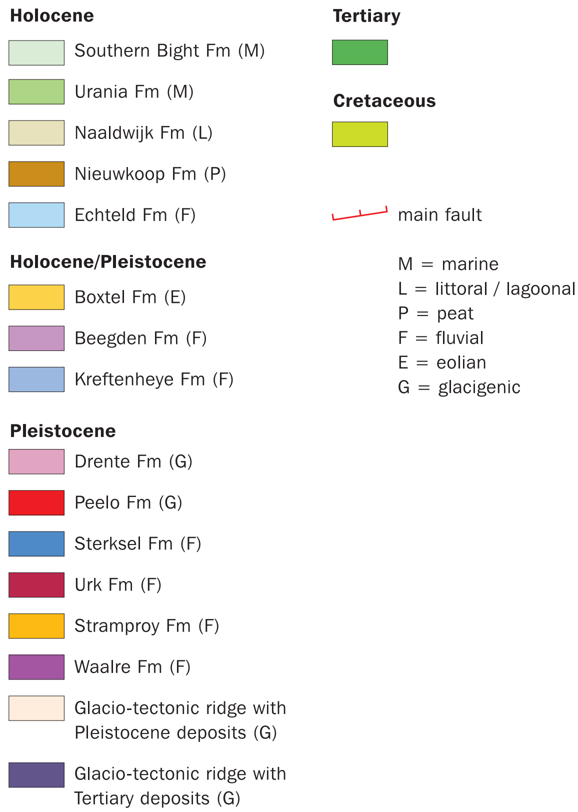


Fig. 1. Geological map of the Netherlands onshore and offshore south of latitude N 54°30′ (data courtesy TNO). The sea bottom north of N 54°30′ consists of the Holocene Southern Bight and Urania formations. See Fig. 3 for stratigraphy. The few outcrops of rocks older than Cretaceous are not shown because of their limited size.

Local insight into the deeper subsoil was already available in the early 17th century, when a water well was drilled in the centre of Amsterdam to a depth of over 60 m, and the recovered sediments were described. No attention was given, however, to the find of marine shells between 25 and 45 m. In 1850, Harting studied the deposits from a water well in Gorinchem (province of Zuid-Holland), and concluded from the presence at 182 m depth of marine shells, which normally live in shallow water, that there must have been a substantial subsidence (Van der Woud, 1998).

In the course of the 20th century the knowledge of the Quaternary improved because of the introduction of new research techniques like pollen and heavy-mineral analyses, radiocarbon dating, paleo-magnetism, amino-acid technology and optically stimulated luminescence. Also the drilling technique improved and thousands of new data points became available. The introduction and adaptation of new ideas can be traced in earlier comprehensive works dealing with the Quaternary of the Netherlands: Faber (1926, 1960), Van der Vlerk & Florschütz (1950), Pannekoek (1956), De Jong (1967), Zonneveld (1971),

Alpine chronology Penck & Brückner (1901-1909)	Geological Map (1942)	Van der Vlerk & Florschütz (1951)	Zagwijn (1957-1989)
Würm	Lower terrace II 8 Loess II 5 & 7	Tubantian	Weichselian
Riss-Würm	Eem deposits II 8'	Eemian	Eemian
Riss	Middle terrace II 6 Fluvio-glacial II 4 Ground moraine II 3 Lower fluvio-glacial II 3' High terrace II 1-2	Drenthian	Saalian
Mindel-Riss		Needian	Holsteinian
Mindel	Pleistocene older than High terrace II 0	Taxandrian	Elsterian Cromerian Bavelian Menapian Waalian Eburonian
Günz-Mindel			
Günz			
Donau-Günz	Upper Pliocene III 3	Tiglian (inclusive of Icenian)	Tiglian
Donau		Praetiglian (inclusive of Amstelian)	Pretiglian
Pre-Donau ?			

cold phases · complex phase · Plio-Pleistocene boundary

Fig. 2. Pleistocene stratigraphy after Penck & Brückner, Geologische Stichting, Van der Vlerk & Florschütz and Zagwijn (partly from Pannekoek, 1956).

Zagwijn (1975, 1989), Van Staalduinen et al. (1979), Berendsen (1996) and De Mulder et al. (2003).

Insights into the Quaternary geology of the Dutch sector of the North Sea developed much later with the mapping of the area by the Rijks Geologische Dienst (later Netherlands Institute of Applied Geoscience TNO – *National Geological Survey*: TNO-NITG). The first studies were published by Oele et al. (1979). In contrast to the onshore, where hundreds of thousands of borehole data are available, the knowledge of the deeper Quaternary in the North Sea is mainly based on geophysical data (Oele et al., 1979; Laban, 1995).

Chronostratigraphy

In the course of time, the chronological subdivision of the Dilivium was greatly altered and adapted to new insights. Nevertheless, it took quite a while before the names Diluvium and Alluvium were gradually replaced by Pleistocene and Holocene. The first subdivisions of the Diluvium were based on faunal variations. For instance, Rutten (1909) distinguished three phases based on the finds of remnants of elephants and rhinoceroses.

The Geologische Stichting (1942) used the Alpine nomenclature of glaciations as described by Penck & Brückner (1901–1909): Günz, Mindel, Riss and Würm. Van der Vlerk & Kuenen (1941) and Van der Vlerk & Florschütz (1950, 1951) presented a local subdivision of the Pleistocene, based on floral and faunal variations (Fig. 2).

Zagwijn (1957, 1961, 1963) detailed the concept of Van der Vlerk & Florschütz and presented a subdivision in which geochronological units are correlated with climatic variations, based on palynological research. He subdivided the Pleistocene into glacial, tree-less periods and interglacial periods with tree growth. Short warmer spell

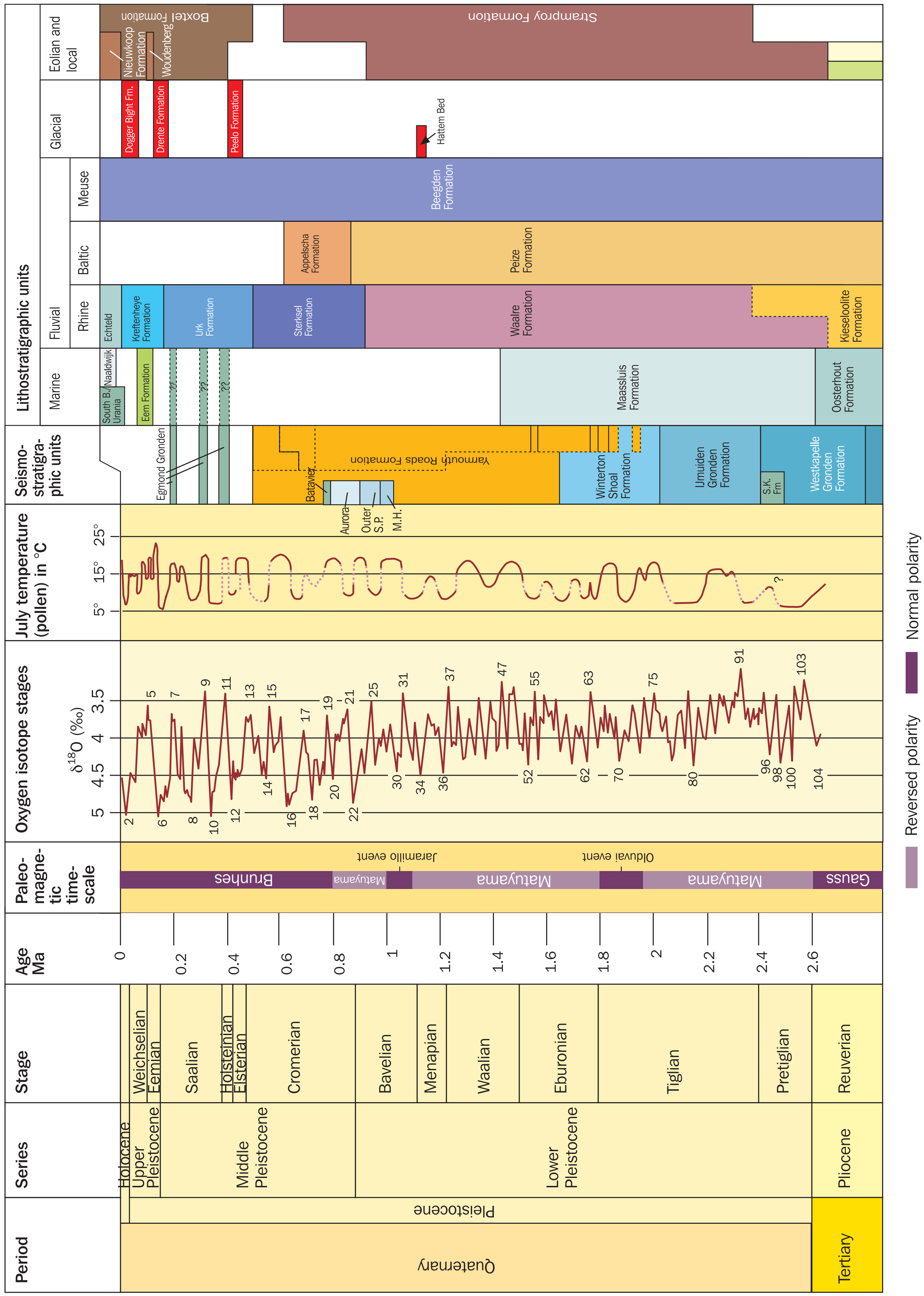
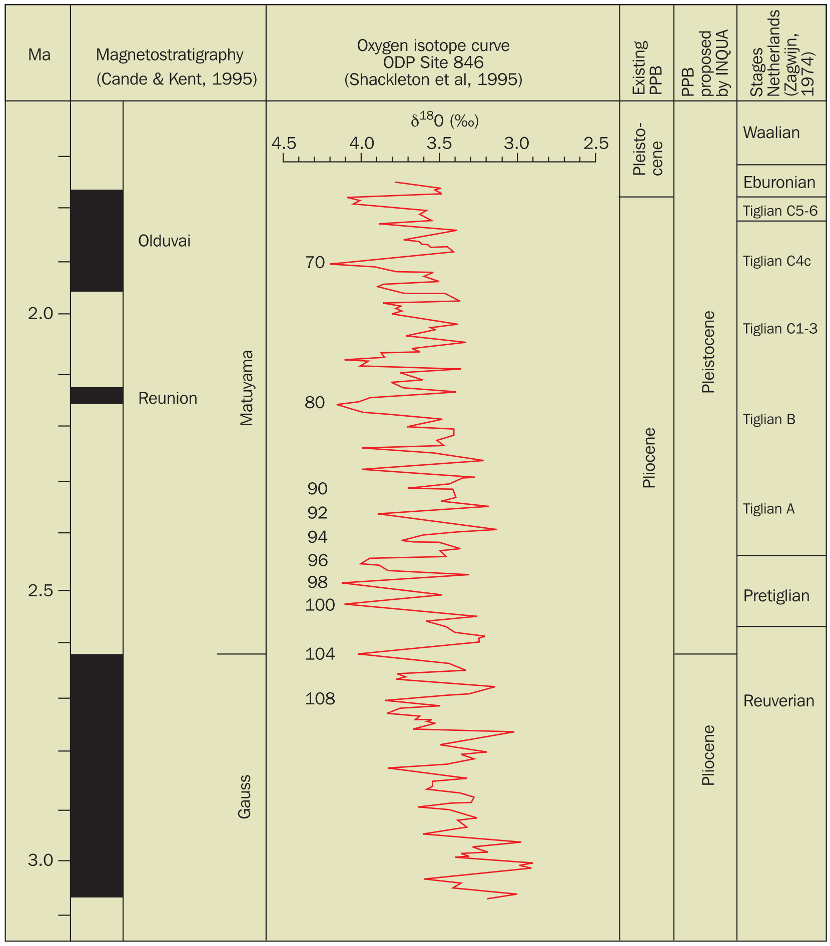


Fig. 3. Pleistocene stratigraphy of the Netherlands (after De Mulder et al., 2003).



PPB: Plio-Pleistocene Boundary

Fig. 4. The possible positions of the Plio-Pleistocene boundary (from: Van Kolfschoten & Gibbard, 1998). In the Netherlands, base Pleistocene is generally taken at base Pretiglian.

within a glacial period were named 'interstadials', and cold spells in interglacial periods 'stadials'. However, the definition of glacial and interglacial periods on the basis of pollen data continues to be a subject of discussion (Turner, 2000). Overall, Zagwijn's subdivision is still valid and is followed here (Fig. 3).

The subdivision into Early, Middle and Late Pleistocene is mainly based on major changes in flora. The boundary between Early and Middle Pleistocene is placed at the beginning of the Cromerian, as in this period characteristic Tertiary trees (*Pterocarya, Carya, Eucommia* and *Tsuga*) are not present anymore (Zagwijn, 1957; De Jong, 1967). The boundary between the Middle and Late Pleistocene is less clear on palynological data. At present it is placed at the base of the Eemian Stage (Van Kolfschoten & Gibbard, 2000).

During the 1948 International Geological Congress it was decided that the Pliocene-Pleistocene, i.e. Tertiary-Quaternary boundary should be placed where the first phenomena indicating a cold phase appear. Consistent with this, Zagwijn (1974) placed the beginning of the Pleistocene at the start of the Pretiglian on palynological data (Figs 2–4). In contrast, many others place this beginning at the end of the magnetic Olduvai event, dated at about 1.8 Ma, i.e. at the boundary Tiglian-Eburonian. In this view, the Pretiglian and Tiglian are Pliocene. It has been proposed to place the Pliocene-Pleistocene boundary at the end of the magnetic Gauss event, i.e. near the end of the Reuverian (Van Kolfschoten & Gibbard, 1998; Zagwijn, 1998). Gradstein et al. (2004) discuss the chronostratigraphic status of the Quaternary in detail.

From the end of the 20th century onwards, the chronological subdivision of the Pleistocene became also based on $^{18}O/^{16}O$ data, either from deep-sea cores (Shackleton & Opdyke, 1977; Shackleton et al., 1984; Shackleton, 1997) or from Greenland and Antarctic ice cores (Dansgaard, 1982; Dansgaard et al., 1993). Gradually, the oxygen-isotope stages are being adopted in the Netherlands, although their correlation with the palynological records is sometimes problematic (Figs 3, 4).

Concerning the numerical ages of the Quaternary, it is surprising that already in the mid-19th century Staring estimated the period in which mammoths were roaming around to be many thousands of years ago. Faber (1926) dated the beginning of the Holocene, based on dendrochronology and sedimentation rates, at 8000 years BP, and the beginning of the Pleistocene at 1 000 000 years, which are, for the early 20th century, good approximations. Many and accurate data are available from the later Pleistocene (Middle and Late Weichselian) and the Holocene. Yet there still are discussions about for instance the time span of the Eemian. Muller (1974), on the basis of counting and extrapolating rhythmites, assumes a duration of 11.5 ka whereas Kukla et al. (1997) suggest a period twice as long.

Lithostratigraphy

The first comprehensive lithostratigraphic classification of the Quaternary was introduced by the Geologische Stichting (1942) for the geological map to the scale 1:50 000. The described units had partly also a geomorphological and chronostratigraphic meaning. For instance, fluvial deposits were subdivided into 'Laagterras afzettingen' (low-terrace deposits) which were correlated with the Würm, 'Middenterras afzettingen' (middle-terrace deposits) which were correlated with the late Riss, and 'Hoogterras afzettingen' (high-terrace deposits), which correlated with the early Riss (Fig. 2). In the end this system and terminology turned out to be unsatisfactory, as for instance in the subsiding North Sea Basin, 'high-terrace' deposits were found in a lower position than 'low-terrace' deposits.

Later attempts at lithostratigraphic classification were based on heavy-mineral analyses (Zonneveld, 1947, 1958). A new stratigraphic framework was established for a next generation of the 1:50 000 geological map (Zagwijn & Van Staalduinen, 1975). This system was a mixture of biostratigraphic analyses, chronostratigraphic interpretations, field relationships, genetic considerations and mineralogical studies, because a purely lithostratigraphic subdivision of the Quaternary was considered unsatisfactory.

A new lithostratigraphy, covering both the onshore and the offshore, has been introduced by Weerts et al. (2003) and De Mulder et al. (2003). This stratigraphy recognizes units from group down to bed level (Figs 1, 3). In this chapter, emphasis will be on the onshore formations. A stratigraphic framework correlating and integrating these formations with those in the offshore was published whilst the present chapter was being finalized (Rijsdijk et al., 2005).

Tectonics, subsidence, climate and sea level

The Netherlands is situated at the southern fringe of the Cenozoic North Sea Basin. The centre of deposition is in the present North Sea (Zagwijn, 1989). The thickness of the Quaternary can be up to 600 m onshore, and up to

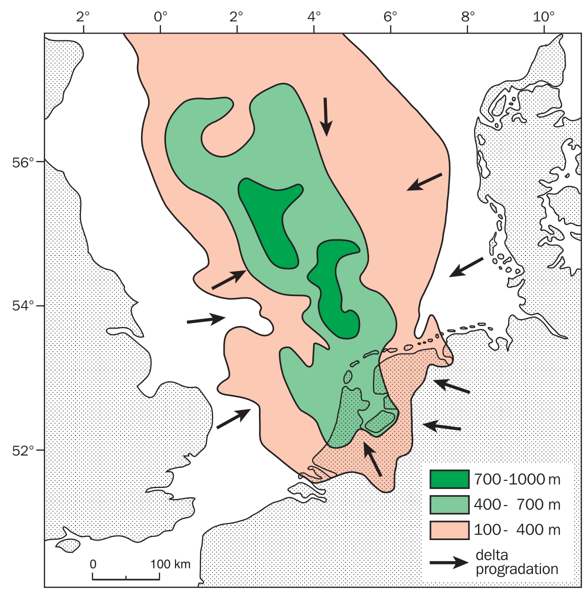
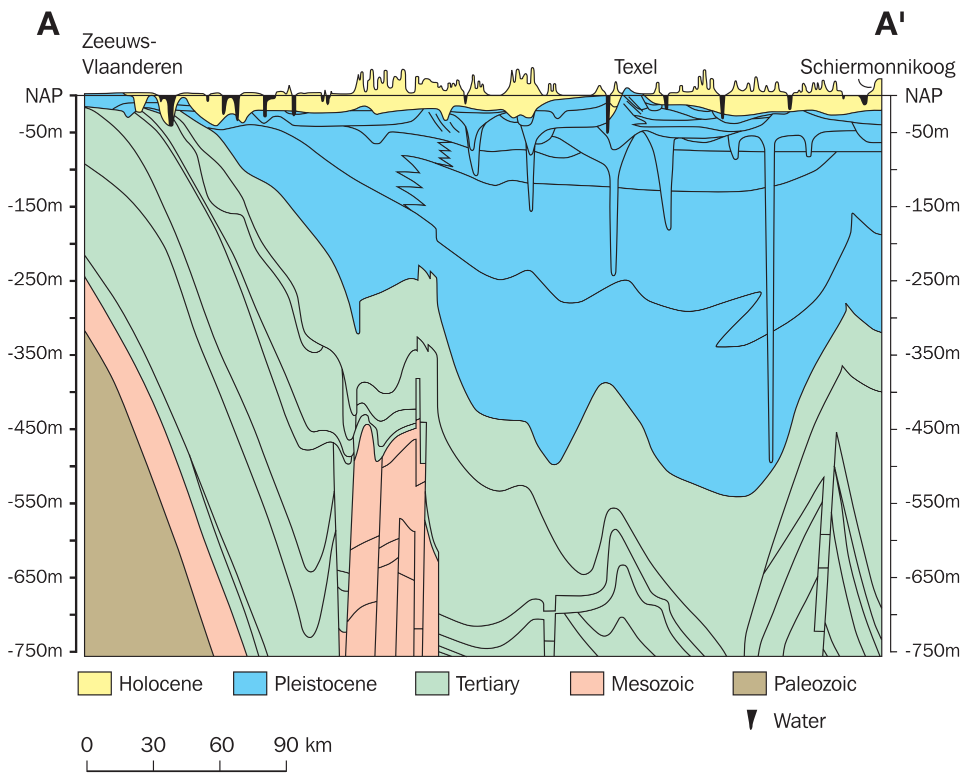


Fig. 5. Thickness of the Quaternary in the south of the North Sea Basin (Laban, 1999).

850 m offshore (Fig. 5; Jeffery et al., 1991). In the basin, fault systems caused differential vertical movements during the Quaternary. Their activity is most conspicuous in the south-eastern Netherlands. In the late Quaternary they influenced river systems (Berendsen & Stouthamer, 2000). However, the exact location of faults in the unconsolidated Quaternary deposits is often hard to establish, notably in the central-western part of the country.

During the Quaternary the sedimentation rate in the basin increased to about ten times the rate prevailing in the Late Tertiary (Zagwijn, 1989), and a varied sequence of fluvial, marine to lagoonal, glacigenic, eolian and periglacial sediments was deposited (Fig. 6). The main cause for the depositional variations is the alternation of many glacial and interglacial periods (Fig. 3: temperature curve). Regressive and transgressive shifts of the coastline occurred with a vertical range of up to 120 m (Fig. 7). During interglacial periods and high sea levels, marine deposits and wedges of deltaic, coastal and lagoonal deposits were formed. In glacial periods, with a low sea level and a coastline situated far to the north, a dry southern North Sea was created, and land-ice masses of northern provenance covered the northern part of the Netherlands several times, giving rise to major changes in geomorphology. These influenced the marine deposition in the following interglacial period. Periglacial deposits such as eolian dunes and periglacial phenomena like cryoturbations and ice-wedge casts also originated.

Fluvial deposition with an eastern, south-eastern or southern provenance occurred during both glacial and interglacial periods. In interglacial periods with a high sea level, fluvial deposition was impeded and anastomosing as



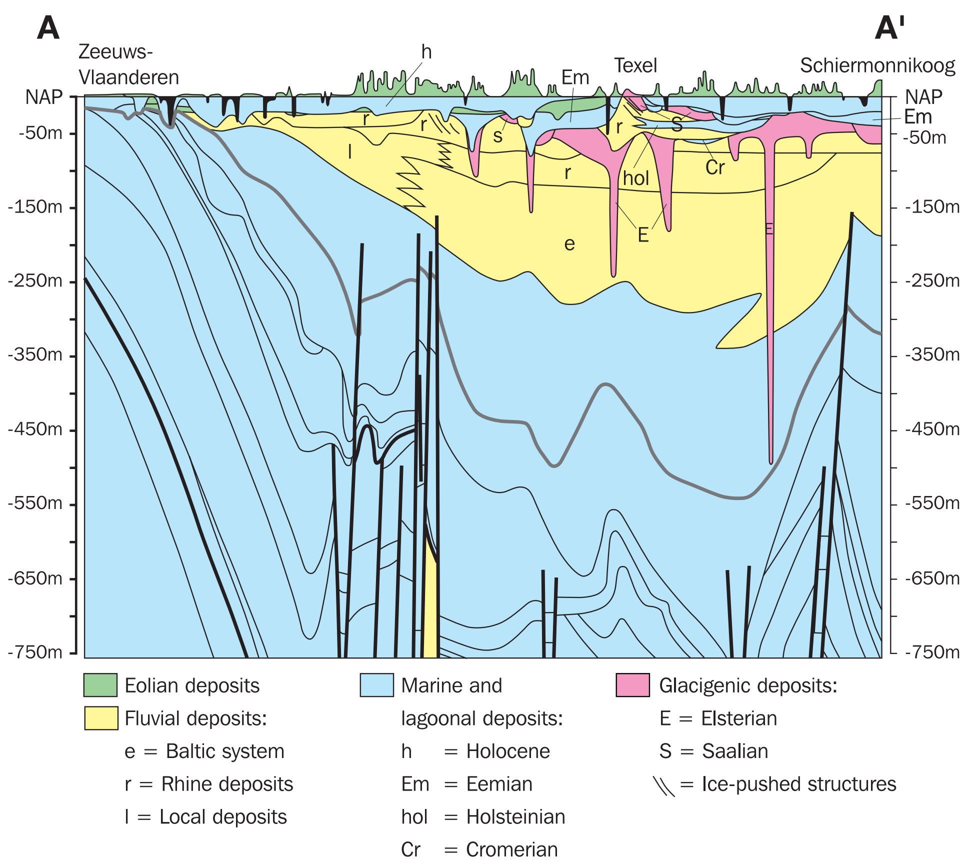


Fig. 6. Age and depositional environment of the Quaternary in a cross section along the coast of the Netherlands. Location in Fig. 8a (after De Gans & Van Gijssel, 1996).

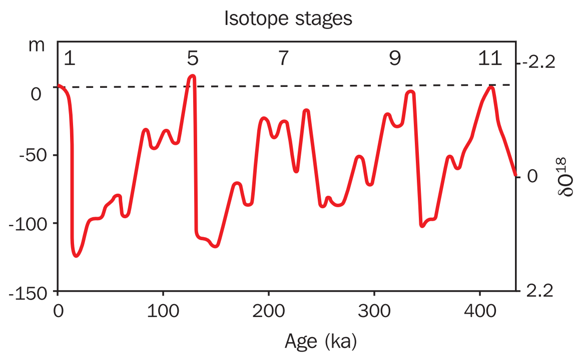


Fig. 7. Global sea-level changes over the past 450 ka (Labeyrie et al., 2002).

well as meandering river deposits were formed. In glacial periods, wide river valleys were formed and subsequently filled by predominantly braided river systems. River-valley shifts were forced by glaciations, subsidence and glacial rebound (De Gans & De Groot, 1995; Törnqvist et al., 2000).

The climatic variations also caused variations in flora and fauna. Given the relatively short time spans these show many extremes. The main floral variation is the disappearance of trees and the increase of grasses and herbs in glacial periods; at times, polar-desert conditions with hardly any vegetation may have occurred. Faunas include subtropical species like rhinoceros (*Decirorhinus kirchbergensis*), monkey (*Macaca florentina*), porcupine (*Hystrix etrusca*), zebra (*Equus robustus*) and bear (*Ursus etruscus*) in the Early Pleistocene interglacial periods, and cold-climate animals like mammoth (*Mammontheus primigenius*), rhinoceros (*Coelodonta antiquitatis*), giant deer (*Cervus giganteus*) and elk (*Alces alces*) in the later glacial periods (Van der Vlerk & Florschütz, 1950; De Jong, 1967).

Pleistocene

Because of the large number of climatic and related sea-level variations, together with the continuous subsidence, the Pleistocene is composed of alternating sequences of glacigenic, marine or lagoonal, fluvial, and eolian deposits (Figs 3, 6). The shifts of coastlines, river systems and ice margins are presented in the paleogeographic maps of Figure 8. For reasons of convenience the Pleistocene stratigraphy is discussed by focussing successively on glacigenic, marine and lagoonal, fluvial, and eolian deposition.

Glacigenic deposits

Glacigenic deposits, sedimentary structures and landforms from five periods of glaciation characterize the Pleistocene (Figs 3, 8d, f, h).

The oldest sedimentary indication for a glaciation is known as the 'Hattem Bed'. This bed is composed of partly highly weathered erratics and gravels of Fenno-scandian provenance, which are interpreted as ice-rafted stones (Lüttig & Maarleveld, 1962; Zandstra, 1971). They are interbedded in *Menapian* fluvial deposits with an eastern provenance (Fig. 3; upper part of Peize Fm). Outcrops of the Hattem Bed did occur in sand pits in glacio-tectonic ridges in the northern Veluwe area near Hattem (Gelderland). The bed was described also from boreholes in Drenthe (Bosch, 1990a, b) and Noord-Holland (De Gans, 1991). In the latter case they are situated at depths of 90 to 100 m. Surprisingly, near Purmerend in Noord-Holland, Fenno-scandian stones and gravel are also found much deeper, between 147 and 163 m depth. Their age and origin are still questionable.

A second indication for the vicinity of land-ice is the Weerdinge Bed described by Ruegg & Zandstra (1977) and by Bosch (1990a, b) from Drenthe. This sand bed is interpreted as a fluvio-glacial deposit. It is dated as *Cromerian* and belongs to the Appelscha Formation.

The third indication corresponds with the first glaciation in which land-ice masses really covered part of the Netherlands and the North Sea area during the *Elsterian* (Figs 3, 6, 8d). Melt-water clays ('potklei' in Dutch, with over 80% lutum) and 'dirty' fine-grained melt-water sands crop out in Drenthe. They are known as the Peelo Formation. Similar deposits occur onshore and offshore in up to 500-m-deep glacial valleys ('tunneldalen' or 'troughs'). Huüse & Lykke-Andersen (2000) suggest that the origin of these valleys can be attributed to a combination of steady-state drainage of melt water below the ice-cap, catastrophic outbursts of melt water, and glacial erosion. This, however, is still questionable. No landforms or tills are known which relate to this glaciation.

According to M.W. van den Berg (personal communication) at least two, or possibly three Elsterian-type glaciations with 'potklei' deposition, dated as isotope stages 8, 12 and 16, took place in the northern Netherlands. This view, however, is not yet generally accepted.

Saalian deposits constitute the fourth indication of a glaciation (Figs 3, 6, 8f). The history of this glaciation has been described for the onshore by Ter Wee (1962), Maarleveld (1953, 1962), Rappol (1984) and Van den Berg & Beets (1987), and for the offshore by Joon et al. (1990) and Laban (1995). During the Saalian, the western and north-western part of the Dutch sector of the North Sea remained ice-free.

This glaciation is characterized by clear geomorphological expressions such as ice-pushed (glacio-tectonic) ridges which partly surround glacial basins, and by deposits like till, fluvio-glacial and mass-flow sands, and lacustro-glacial clay. Surprisingly, no deep glacial valleys ('tunneldalen') are known. The glacio-tectonic ridges were formed by piping and pushing around the southern ends of glacier tongues. They consist of tilted and faulted strata composed of the material available: predominantly Rhine sands in the western ridges, Rhine sands and sands of eastern provenance in the central ridges, and the latter together with Tertiary strata in the eastern ridges. The largest glacio-tectonic ridge, the East Veluwe, is now up to 100 m high, about 50 km long and up to about 12 km wide. The glacio-tectonic ridges are flanked at their southern sides by outwash plains ('sandrs') which are composed of fine to coarse sand derived from the ice-pushed ridges themselves together with some fluvio-glacial material (Maarleveld, 1953, 1962; Ruegg, 1983). In places, overflow channels occur in the ridges (Berendsen, 1996).

The glacial basins mainly occur in the central Netherlands (Fig. 8f); however, Joon et al. (1990) describe such basins from the offshore area as well. The basins are filled with lacustro-glacial clays which show varves in their bottom layers (Jelgersma & Breeuwer, 1975). Part of these clays is thought to have a fluvial origin (De Gans et al., 2000). Based on the deduced lake level in the Amsterdam Glacial Basin (Fig. 9), De Gans et al. (2000) suggest that a large lake covered central Noord-Holland and the IJsselmeer area during the Late Saalian.

Tills, with thicknesses up to 10 m, occur in the subsoil of the northern Netherlands and the adjacent North Sea. They crop out in Drenthe and Friesland, and are mainly composed of silts and sands, and of numerous crystalline and other erratics of a Fenno-scandian provenance (Zandstra, 1983). From the island of Texel, eastwards via Wieringen and Gaasterland to Steenwijk in Overijssel, a series of small hills (+10 to +20 m NAP; NAP = Normaal Amsterdams Peil = Dutch ordnance datum = approximately sea level) occurs, covered by till and composed of till and tilted eolian and fluvial deposits (Fig. 7f: till-cored ridges). Glacial basins with lacustro-glacial deposits related to these hills are scarce. The hills are thought to represent older, overridden glacio-tectonic ridges (Rappol, 1984). The Saalian glacigenic deposits are referred to as Drente Formation.

Weichselian fluvio-glacial sands, glacio-marine clastics and subglacial valleys in the north-westernmost part of the Dutch North Sea sector represent the fifth glacial series (Figs 3, 8h). The sediments occur at shallow depth and are referred to as Dogger Bight Formation. The related land-ice masses had their origin in England and Scotland (Laban, 1995).

Marine and lagoonal deposits

During the Tertiary-Quaternary transition, most of the Netherlands was a shallow sea (Fig. 8a). In the marine sediments this transition is represented by the boundary between the Oosterhout and Maassluis formations. The Lower Pleistocene shell-bearing sands and clays of the Maassluis Formation were deposited in marine, tidal and littoral environments by prograding deltas (Zagwijn & Van Staalduinen, 1975; Kasse, 1988). Their thickness may reach 300 m below the islands of Vlieland and Terschelling (Fig. 6). Jansen et al. (2004) argue that the lower part of the Maassluis Formation is laterally equivalent to the shallow-marine Oosterhout Formation and gets progressively younger towards the west. The upper part is laterally equivalent to the Peize Formation (Fig. 3).

Marine Waalian sediments are described by Zagwijn (1975) and Kasse (1988) from the south-western Netherlands and the southern North Sea.

Middle Pleistocene sands with marine shells from the same area attain a thickness of up to 10 m. Most of them are considered now to be fluvial deposits containing reworked marine shells from earlier marine interglacial deposits (De Gans & De Groot, 1995; De Gans & Wassing, 2000; Törnqvist et al., 2000).

In the northern Netherlands and the southern North Sea, shell-bearing sands and clays are regarded as marine deposits in situ, dating from various intervals in the Cromerian Complex and the Holsteinian (Figs 8c, e). In the northern Netherlands the stratigraphic position of these deposits is evident as they occur below glacigenic Saalian or in between glacigenic Saalian and Elsterian deposits (Fig. 6).

The Late Pleistocene Eemian sediments of the Eem Formation are the best known marine Pleistocene deposits (Figs 3, 6, 8g). However, like the earlier Cromerian and Holsteinian marine deposits, part of the marine Eemian sands in the Zuid-Holland area and elsewhere are considered to be fluvially reworked (De Gans & De Groot, 1995; De Gans & Wassing, 2000; Törnqvist et al., 2000).

Basal peat beds, like those found over large areas at the base of the Holocene, are absent in the Eemian (De Gans et al., 2000). In contrast, the lowermost Eemian in the deep glacial basins comprises a black layer of sapropel and diatomaceous earth, informally called 'Harting layer' (Fig. 9). The base of the overlying lagoonal Eemian clays is in the Amsterdam Basin at about –80 m NAP; the top is in places, at the fringe of the basin, at –10 m NAP (De Gans et al., 2000; De Gans & Wassing, 2000). In the central Holland area, outside the glacial basins, the top of the Eemian marine deposits is due to subsidence and erosion generally not above –20 m NAP. These deposits are composed of a complex alternation of sand and clay. In the southern North Sea area, extensive lagoonal clayey deposits were laid down towards the end of the Eemian (Cameron et al., 1989).

Fluvial deposits

Fluvial deposition of sands and clays already took place from the beginning of the Pleistocene in the south-eastern

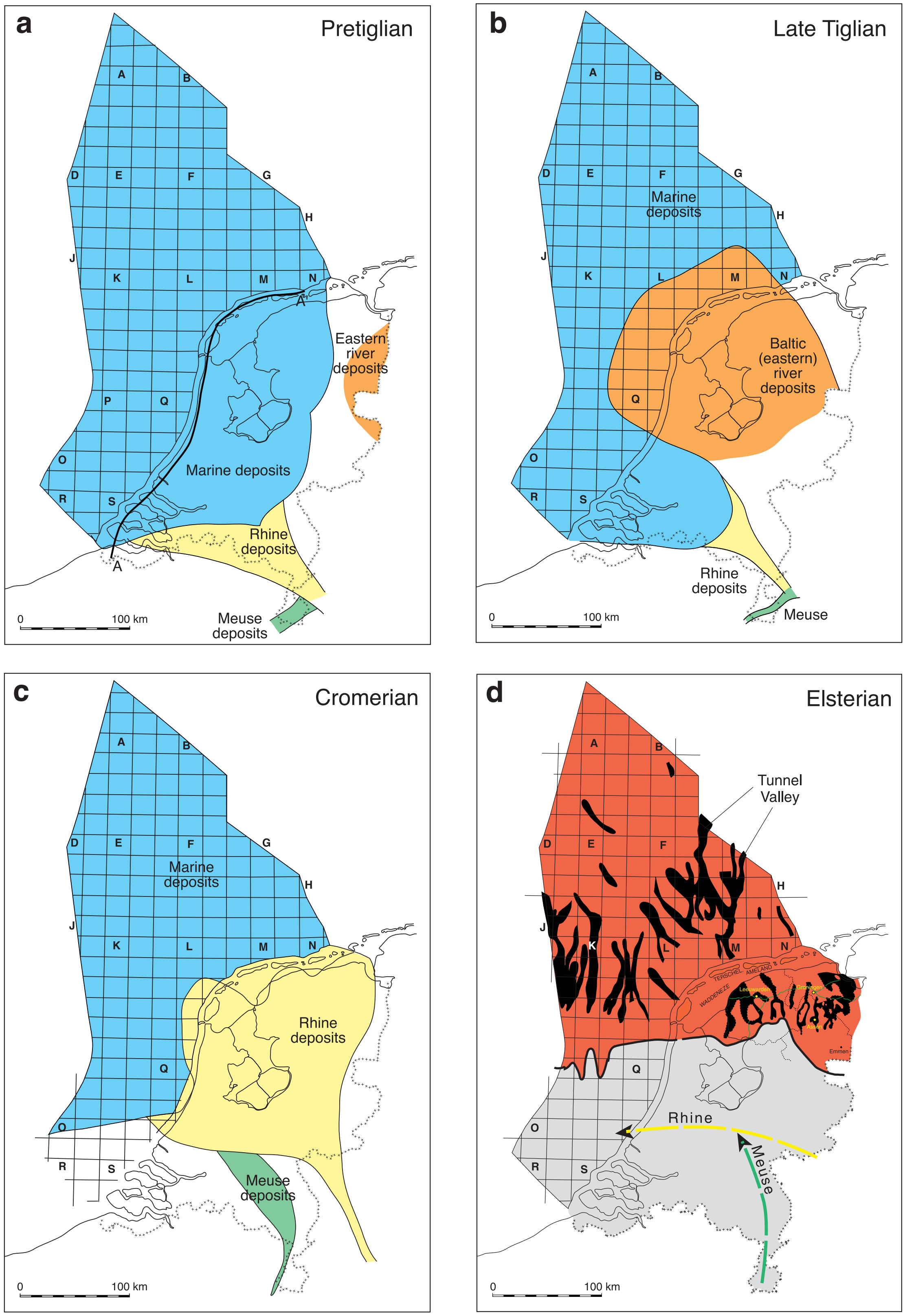
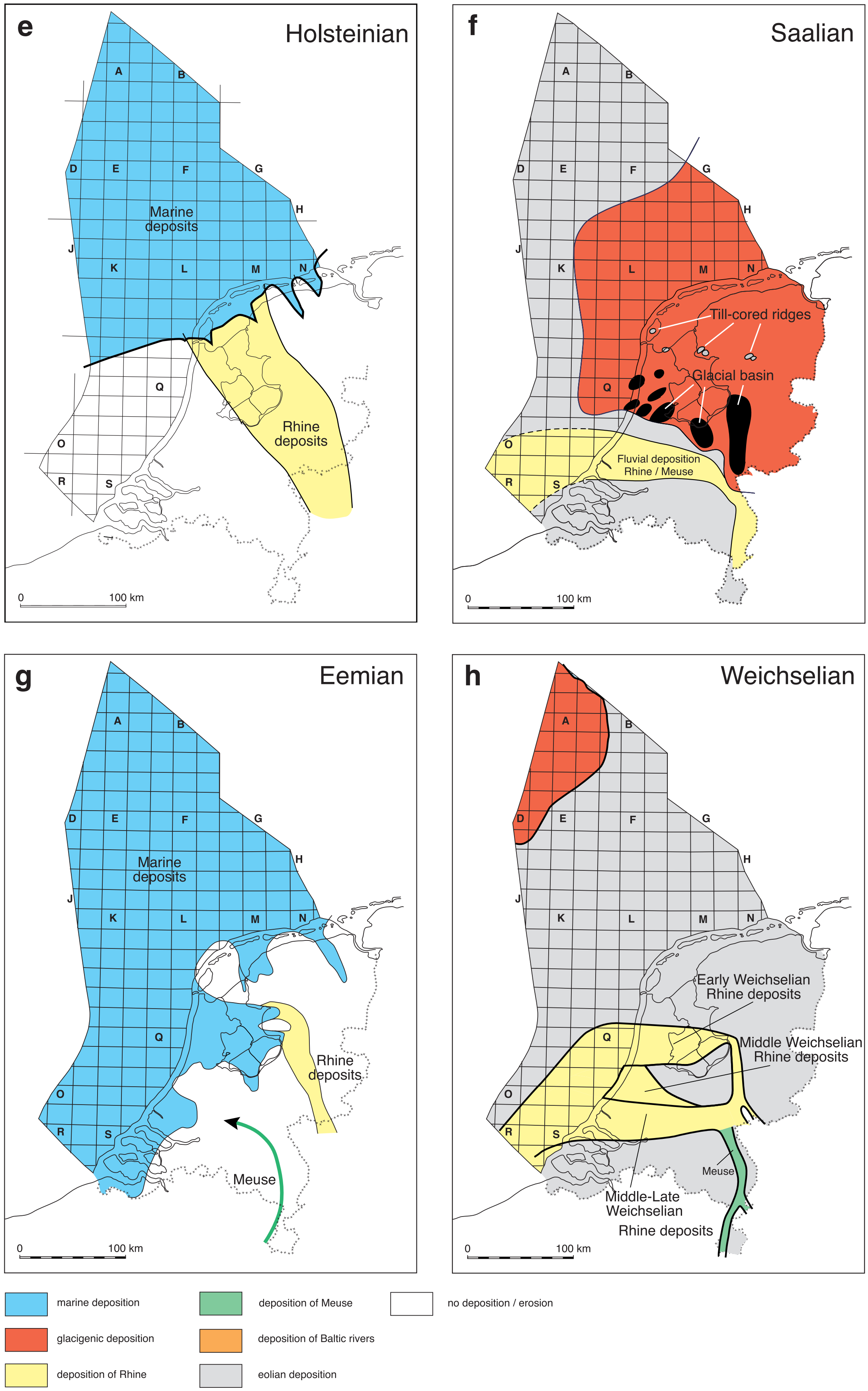


Fig. 8. Paleogeographic maps of the Pleistocene in the Netherlands (after Zagwijn, 1975; Van den Berg & Beets, 1987; Laban, 1995; Bosch, 1990a, and Kasse, 1988).


e
Holsteinian
Marine deposits
Rhine deposits
0 100 km
f
Saalian
Till-cored ridges
Glacial basin
Fluvial deposition Rhine / Meuse
0 100 km
g
Eemian
Marine deposits
Rhine deposits
Meuse
0 100 km
h
Weichselian
Early Weichselian Rhine deposits
Middle Weichselian Rhine deposits
Meuse
Middle-Late Weichselian Rhine deposits
0 100 km
marine deposition
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deposition of Rhine
deposition of Meuse
deposition of Baltic rivers
eolian deposition
no deposition / erosion

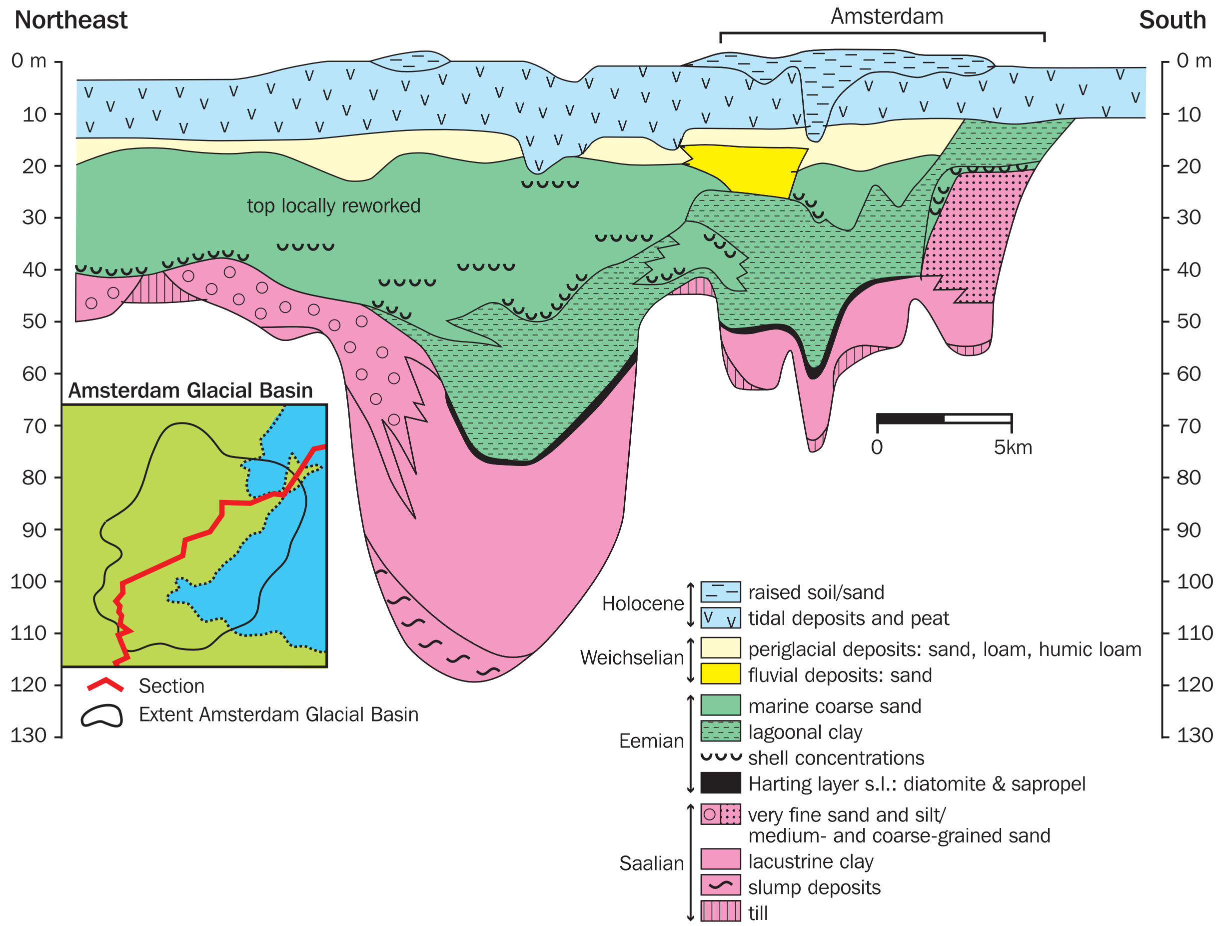

Fig. 9. Section showing the fill of the Amsterdam Glacial Basin (De Gans et al., 2000). The substrate of the basin is composed of pre-Saalian fluvial deposits. Blue on inset location map represents present-day IJsselmeer or Markermeer.

and eastern Netherlands: either from the south-eastern, Rhine and Meuse source or from an eastern Baltic source (Eridanos system; Figs 8a, b).

In the Late Tertiary, the Rhine took a course through the Roer Valley Graben as a result of tectonic activity. Here, the Rhine and Meuse deposited the coarse sands of the Kieseloolite Formation. During the Tiglian the deposition of the Rhine varied in accordance with the shifts of coastlines (Zagwijn, 1975). The stratotype of the Tiglian is located near Tegelen, north of the Roer Valley Graben. These fluvial clays, which were used already in Roman times for the production of 'tegulas' (tiles), were deposited by meandering rivers, and contain a vast number of extinct floral and faunal remains. The latter include *Tapirus arvenensis*, *Anancus arvenensis* (mastodont) and *Dicerorhinus etruscus* (Westerhoff et al., 1998). At the end of the Tiglian the fluvial deposition of the Rhine replaced the already mentioned marine environment in the west, forming an extensive delta system (Waalre Fm). On top these deposits a compacted peat or brown-coal layer may be present. As a result of the formation of this layer and of a long period of non-deposition the upper part of the deposits is decalcified.

From the Late Tertiary on, a fluvial 'Baltic' system from the east, recently called the Eridanos river, with a source as far away as Finland, formed extensive deltas (Overeem, 2002). These deposits (Peize and Appelscha formations) are predominantly composed of white quartz sands with no, or very little calcium carbonate. They were formed until the middle of the Cromerian and may attain thicknesses onshore of over 250 m; offshore they are much thicker (Cameron et al., 1993).

In the same period, local rivers from the south, with a provenance in the Brabant Massif, formed deltas of a much smaller scale, composed of both sand and clay. The sands have a greyish colour and a low calcium carbonate content (Stramproy Fm).

During the Cromerian the Rhine became the major supplier of fluvial deposits, forming stacked deltas predominantly composed of reddish sands (Figs 6, 8c; Sterksel Fm).

The Meuse is only of minor importance in this pattern of sediment supply (Beegden Fm). However, due to the uplift of the Ardennes, an extensive flight of fluvial terraces is present along the Meuse, dating from the Late Tertiary to the Late Weichselian and ranging in elevation from +40 to +250 m NAP (Van den Berg, 1996).

From the Elsterian onward the sedimentation patterns of the Rhine can be reconstructed with reasonable accuracy (Urk and Kreftenheye formations; Figs 8d–h). In interglacial periods the valleys have a north-western orientation towards the areas of maximum subsidence in the North Sea Basin. In glacial periods, river courses shifted towards the west and south-west as a result of the extension of land-ice margins and/or as a result of glacial rebound. The last phenomenon is demonstrated clearly during the Weichselian, when the Rhine shifted from a north-western to a western course before bending south towards the English Channel.

Eolian deposits and permafrost indicators

Eolian deposits were formed on an exceptionally large scale during the late Middle and Late Weichselian (Figs 3, 6, 8h). They belong to the Boxtel Formation. They are also referred to as 'cover sands' (dekzanden) as they cover large parts of the older deposits all over the Netherlands. Generally the cover sands are about 2 m thick. In the Roer Valley Graben in Noord-Brabant, they exceed 10 m in thickness, but here eolian deposits from older glacial periods are also present.

The late Middle Weichselian 'Old Cover Sands' are composed of loam and fine sand. Later 'Young Cover Sands' consist of medium-grained sands. They often show longitudinal or parabolic dunes. The eolian sedimentation of cover sands came to an end in the Early Holocene (Cleveringa et al., 1977). The sands were in part derived from the then dry southern North Sea area and from local sources (Schwan, 1988).

Wind-blown dunes were formed also in the floodplain of the Late Weichselian Rhine. These 'rivierduinen' (river dunes) consist of reworked fluvial sands and are much coarser than the Young Cover Sands. Generally these dunes have a SW-NE orientation. Locally their tops crop out in the Holocene floodplain (cf. Fig. 14).

Gravel beds in deflated areas, often contain wind-faceted stones. The main gravel bed, known as the Beuningen Gravel Bed, is dated as late Middle Weichselian (Van der Hammen & Wijmstra, 1971; Kolstrüp, 1980).

The Middle Weichselian permafrost in the Netherlands was presumably about 20 m thick (De Gans, 2000). The intense eolian activity at the end of the Weichselian was possibly related to the disappearance of this permafrost, the deep incision of river channels and the resulting lowering of the groundwater level (Schwan, 1988). Indicators of the Weichselian permafrost are ice-wedge casts and remnants of ice-cored mounds (pingo scars). The major level of ice-wedge casts occurs just below the Beuningen Gravel Bed (Van der Hammen et al., 1967; Kolstrüp, 1980). The pingo scars are scattered all over the northern Netherlands and are related to Late Weichselian valley systems (De Gans, 1983). Ice-wedge casts are also known from older glacial periods; the oldest so far date from the Menapian (De Mulder et al., 2003).

In contrast to the northerly cover sands, loess deposits dating from the Weichselian and Saalian are found in the extreme south of the Netherlands (Zuid-Limburg). Patches of Weichselian loess also occur in the central Netherlands (Edelman, 1950). Eolian sands dating from the Elsterian and Saalian glaciations are mainly found south and west of the lines of maximum extent of these glaciations. They also occur directly underlying the Saalian till in the north of the Netherlands, or intercalated in Elsterian deposits (Figs 8d, f).

Holocene

Stratigraphy

The Holocene chronological subdivision used in the Netherlands is primarily based on palynological data (De Jong, 1982). The beginning of the Holocene, defined as the period of warmer climate after the last glacial period, is characterized by the first occurrence of thermophilous plants and trees in pollen diagrams. The subdivision into Preboreal, Boreal, Atlantic, Subboreal and Subatlantic (Fig. 10) was introduced in the Netherlands by Florschütz (1933) and taken from the work of Blytt & Sernander (Zagwijn, 1986).

The Holocene is now generally accepted to start at 10 000 BP. However, the Geologische Stichting (1942) placed this beginning at 20 000 BP, and included what we now call the Late Glacial (Weichselian) in the Holocene on the basis of the first signs of climatic improvement.

The lithostratigraphy of the Dutch Holocene, originally based on the work of Staring (1854, 1856, 1860), was detailed by the geogenetic soil studies of Edelman (1950) and the work of the Geologische Stichting and the later Rijks Geologische Dienst (Geological Survey). Hageman (1969) postulated the idea of synchronous clastic sedimentation all over the coastal areas, related with transgressive periods, and of peat layers correlated with regressive periods. The older marine, estuarine and lagoonal deposits Calais I-IV were related with the peri-marine fluvial deposits Gorkum I-IV, and the younger marine, estuarine and lagoonal deposits Dunkerque 0-III, were correlated with the fluvial deposits Tiel 0-III. Later this idea of synchronous clastic deposits all over the coastal areas was abandoned and replaced by the notion of synchro-

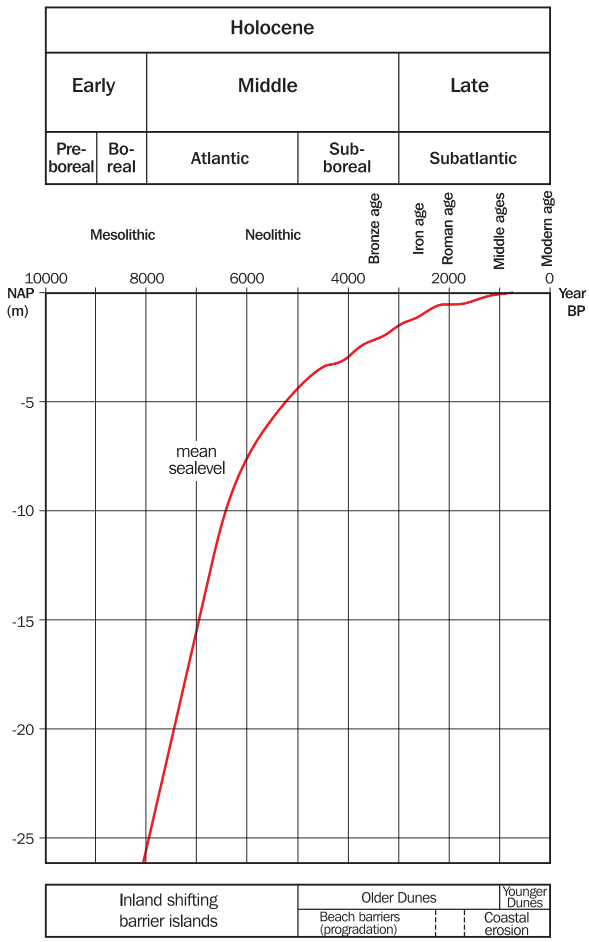

Fig. 10. Holocene chronostratigraphy and sea-level rise in the Netherlands (after Zagwijn et al., 1985).

nous deposition in separate tidal basins (Beets et al., 1992; Van der Spek, 1994). At present the marine Holocene sediments are referred to as Naaldwijk Formation onshore, and as Southern Bight Formation and Urania Formation offshore. Holocene peat is classified as Nieuwkoop Formation, and the fluvial deposits of Rhine and Meuse as the Echteld and Beegden formations respectively (Figs 1, 3).

The main processes influencing the Holocene sedimentation are the rise in sea level, and increases in vegetation density and precipitation. In the Subatlantic period, man gradually becomes a major geological factor.

Sea-level rise

It was already postulated around 1850 that a sea-level rise must have happened in the Alluvial period (Van der Woud, 1998). Early graphs of sea-level rise were published by the Geologische Stichting (1942) and Bennema (1954). Jelgersma (1961) constructed the first reliable sea-level curve for the western Netherlands on the basis of a large number of radiocarbon data. Van der Plassche (1982) slightly modified this curve. The depicted rise of sea level is in fact relative, as subsidence is part of the process. The curve shows two major periods: one before about 5000 BP, when sea-level rise is relatively fast, and one after 5000 BP, when it is slowing down (Fig. 10).

The most recent part of the curve is known from data from Amsterdam, where in 1683/84 benchmark stones were implanted in the then sea wall of the town. Measurements related to this benchmark, allowed sea level to be monitored several times a day. They were carried out until about 1930 (De Gans, 2006).

Marine and lagoonal deposits, and peat

During the coldest part of the Weichselian, around 20 000 BP, sea level was at about –120 m NAP (Fig. 6) and coastlines were situated west of the English Channel and near the Dogger Bank. Due to sea-level rise the northerly coastline moved south-eastward (Fig. 11) and an eastward expanding wedge of Holocene deposits was formed, composed from bottom to top of (basal) peat, lagoonal and tidal deposits, and again peat (Fig. 12).

The basal peat layer is compacted and has a maximum thickness of 2 m. It comprises remnants of wood. It became intensely eroded by tidal channels in the seaward coastal zone and even more so in the North Sea area, where it has a patchy occurrence. The oldest basal peat layer onshore is dated as Boreal and found at depths of –15 to –20 m NAP. Offshore, Jelgersma (1979) dated basal peat layers at depths of –46 to –47 m NAP as earliest Preboreal.

The directly overlying lagoonal clay often contains large amounts of the gastropod *Hydrobia* sp. It is generally less than 2 m thick, and grades upwards into tidal, Atlantic and early Subboreal sands and clays which are related to the relatively fast rise in sea level prior to about 5000 BP. Their sedimentation was largely controlled by the rate of this rise and the supply of sediment as the continuous creation of accommodation space in back-barrier basins by the rising sea level outran the sediment supply (Beets et al., 1996).

The tidal deposits contain several thin (up to 1 m thick) peat layers which indicate fresh-water influxes. These peat layers are probably related to a temporary closing off of the back-barrier basins and are restricted to these basins.

Dated at about 5000 BP, the oldest preserved beach barriers are found in the western Netherlands (Figs 12, 13a, 14). From 5000 to 2000 BP the coastline prograded and series of beach barriers were formed westward of the previous ones. This progradation occurred in spite of a substantially rising sea level (Fig. 10) and is the result of a large supply of sand from the offshore coastal zone (Beets et al., 1992). The barriers have lengths of tens of kilometres and widths up to several hundreds of metres. They were formed to just above mean sea level. On top of the barriers, sands were blown to dunes up to 10 m high, known as the 'Older Dunes'. The arrangement

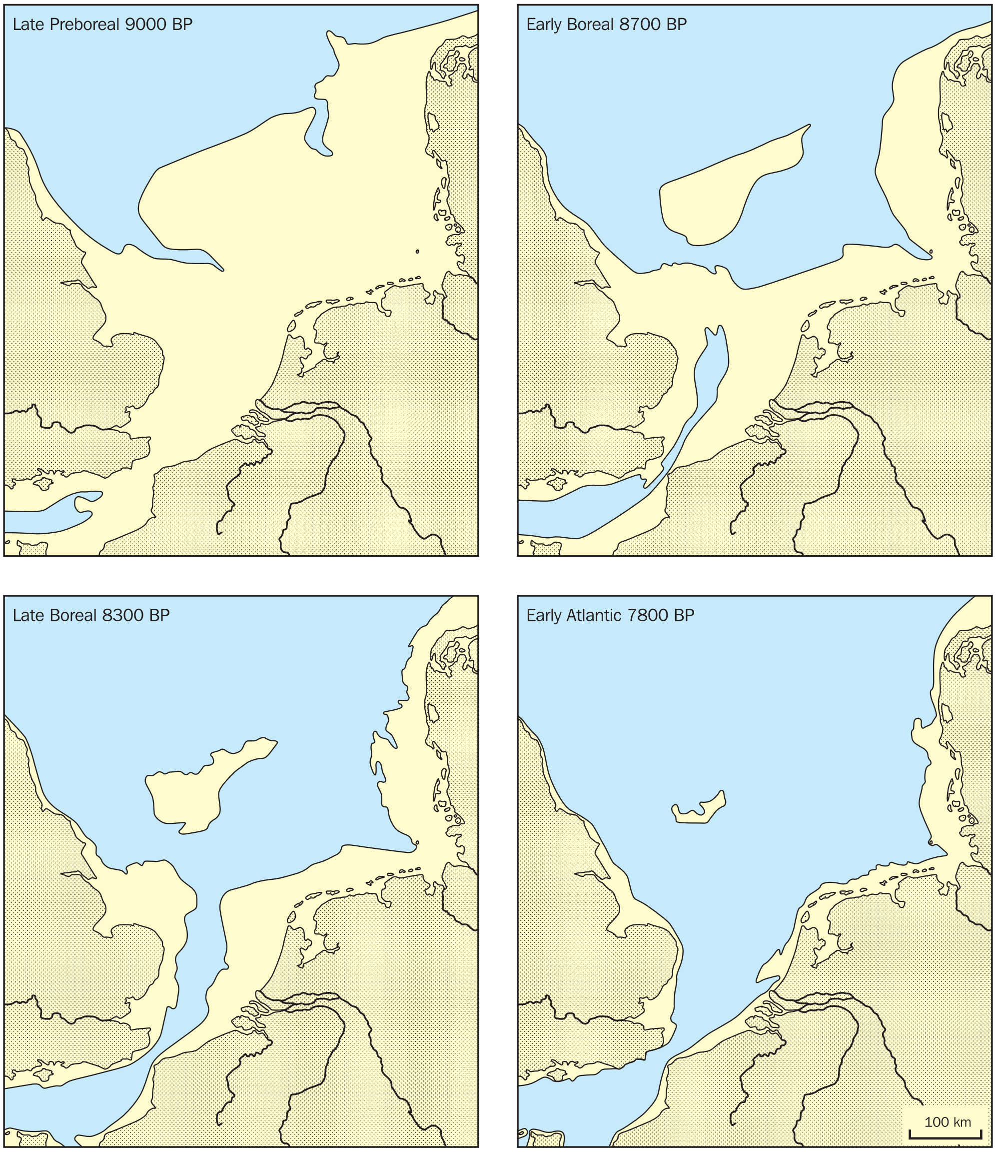

Fig. 11. Paleogeographic maps of the Early Holocene in the south of the North Sea Basin (after Zagwijn, 1986). Yellow: land; blue: sea.

of these barriers suggests the presence of estuaries near Hoek van Holland (Rhine-Meuse system) and the mouth of the Oude Rijn (Old Rhine).

The area east of the just established barriers changed from a tidal flat into a freshwater lagoon at the end of the Atlantic period. In Holland and Zeeland, peat accumulated on top of the tidal deposits. This accumulation ended in the Roman period as a result of draining and cultivation of land. The absence of a barrier system in the northern Netherlands, resulted in continued tidal deposition (Figs 12a, 13b). Due to a sediment deficit the Waddenzee and the Flevomeer (later Zuiderzee, now IJsselmeer) area remained open.

In the later Subatlantic period, tidal deposition increased again in the coastal zone of Zeeland and Holland, and tidal deposits were laid down in estuaries and on top of the peat. In Zeeland large parts of peat lands were eroded (Fig. 13c).

From about 2000 BP onward, progradation of the coast ended and since 1000 AD coastal erosion started. The foreshore became steeper and large quantities of sand were deposited on the coast. This resulted in the formation of the so-called 'Younger Dunes' or coastal dunes. The Younger Dunes in the north-west of Noord-Holland consist, in contrast to those in the south, of sands poor in calcium carbonate. Elsterian ice-pushed ridges in the subsoil, composed of fluvial sands related to the Eridanos fluvio-deltaic system, may have contributed to the provenance of these sands.

Offshore, most of the Holocene sediments belong to

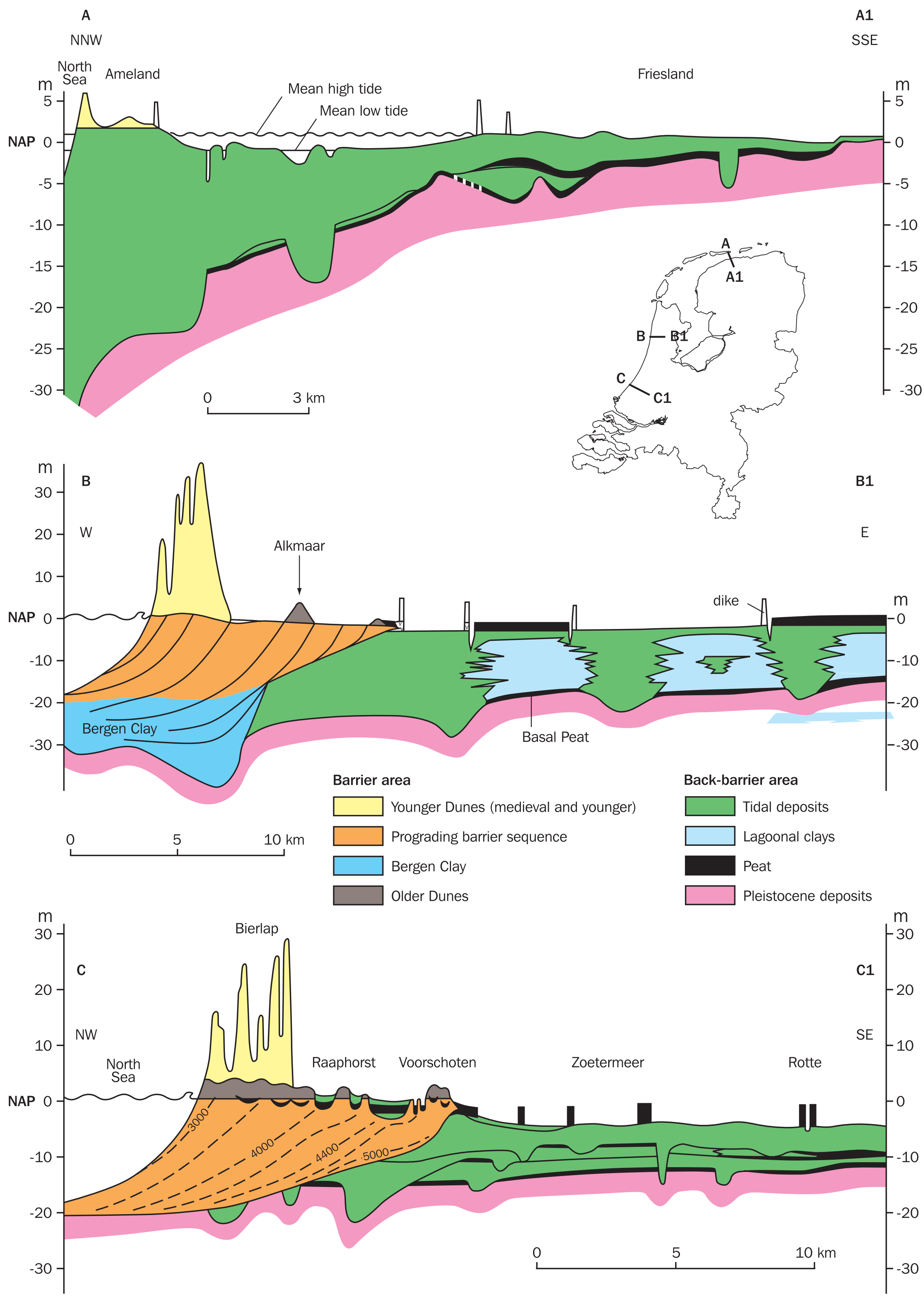

Fig. 12. Cross sections showing the Holocene sequences in the coastal plains of Friesland (A-A1), Noord-Holland (B-B1) and Zuid-Holland (C-C1) (after Van der Valk, 1992, and Van der Spek, 1996). The local absence of peat at the surface in Noord- and Zuid-Holland is the result of erosion and exploitation. 5000, 4000, etc. in the barrier sequence refer to years BP.

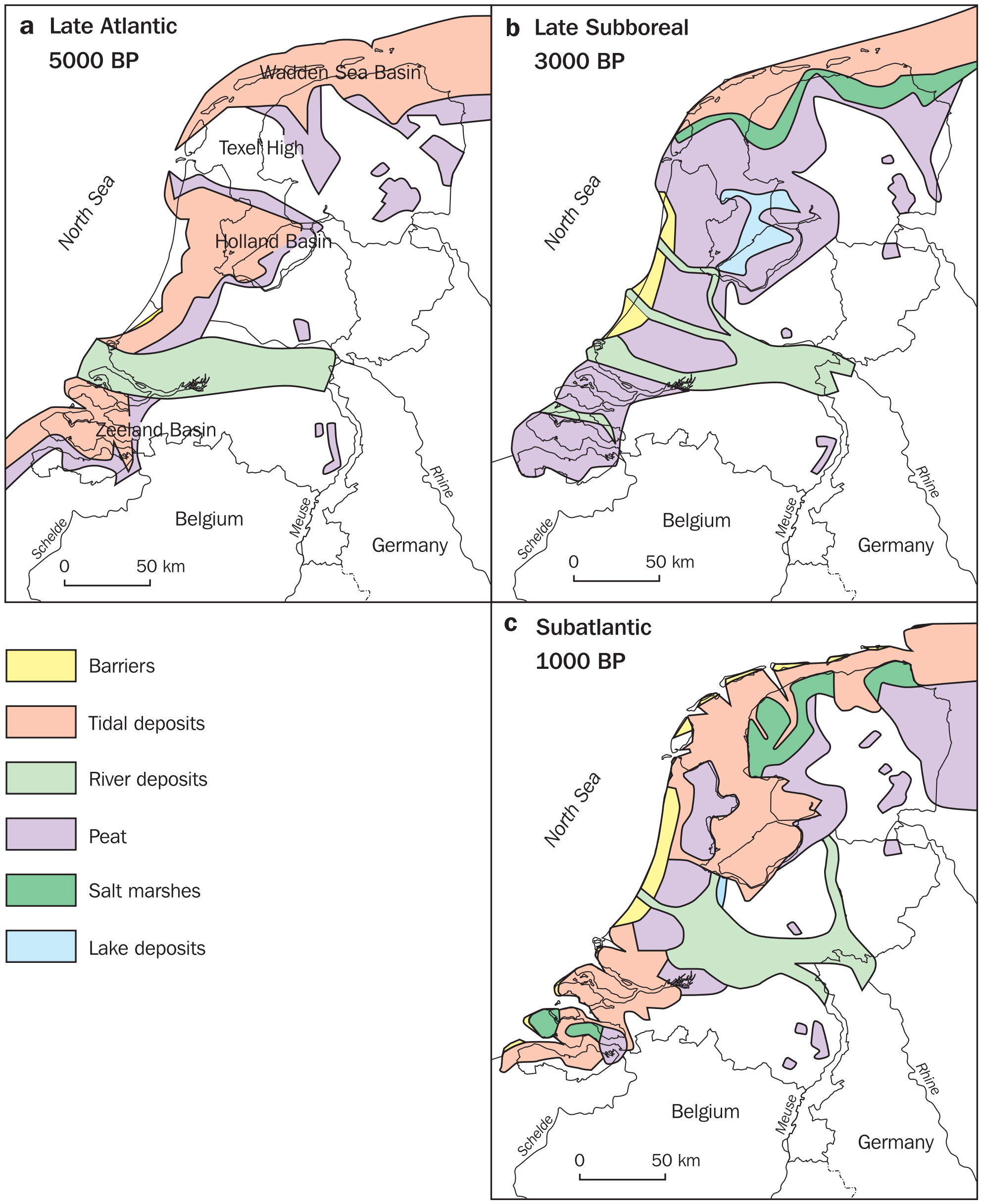

Fig. 13. Paleogeographic maps of the Middle and Late Holocene in the Netherlands (after Beets & Van der Spek, 2000).

the Southern Bight Formation (Figs 1, 3). They consist of shallow-marine sands, with locally reworked glacial material in the north of the area. In the south, sand waves and megaripples occur. Much of the central-northern offshore area is composed of fine sands and clays, known as Urania Formation. The Holocene deposits offshore generally do not exceed 15 m; they may be absent locally.

River and brook deposits

It is generally accepted that at the end of the Weichselian the main river systems changed from braiding into meandering as a result of an increased vegetation cover and a decreased peak discharge and sediment load. Sedimentation patterns changed from broad river plains with river dunes and shallow water depths into narrower plains with meandering channels. Natural levees were formed and extensive overbank clays deposited.

As a result of sea-level rise, stabilized river banks and strong in-channel aggradation, an anastomosed river pattern developed in the coastal zone of the central Netherlands (Fig. 14a; Törnqvist, 1993; De Groot & De Gans, 1996; Makaske, 1998). More to the east the meandering sedimentation by Rhine and Meuse continued with multiple channel shifts and avulsions (Berendsen, 1982; Stouthamer, 1991).

From the end of the Atlantic (about 5000 BP) on-

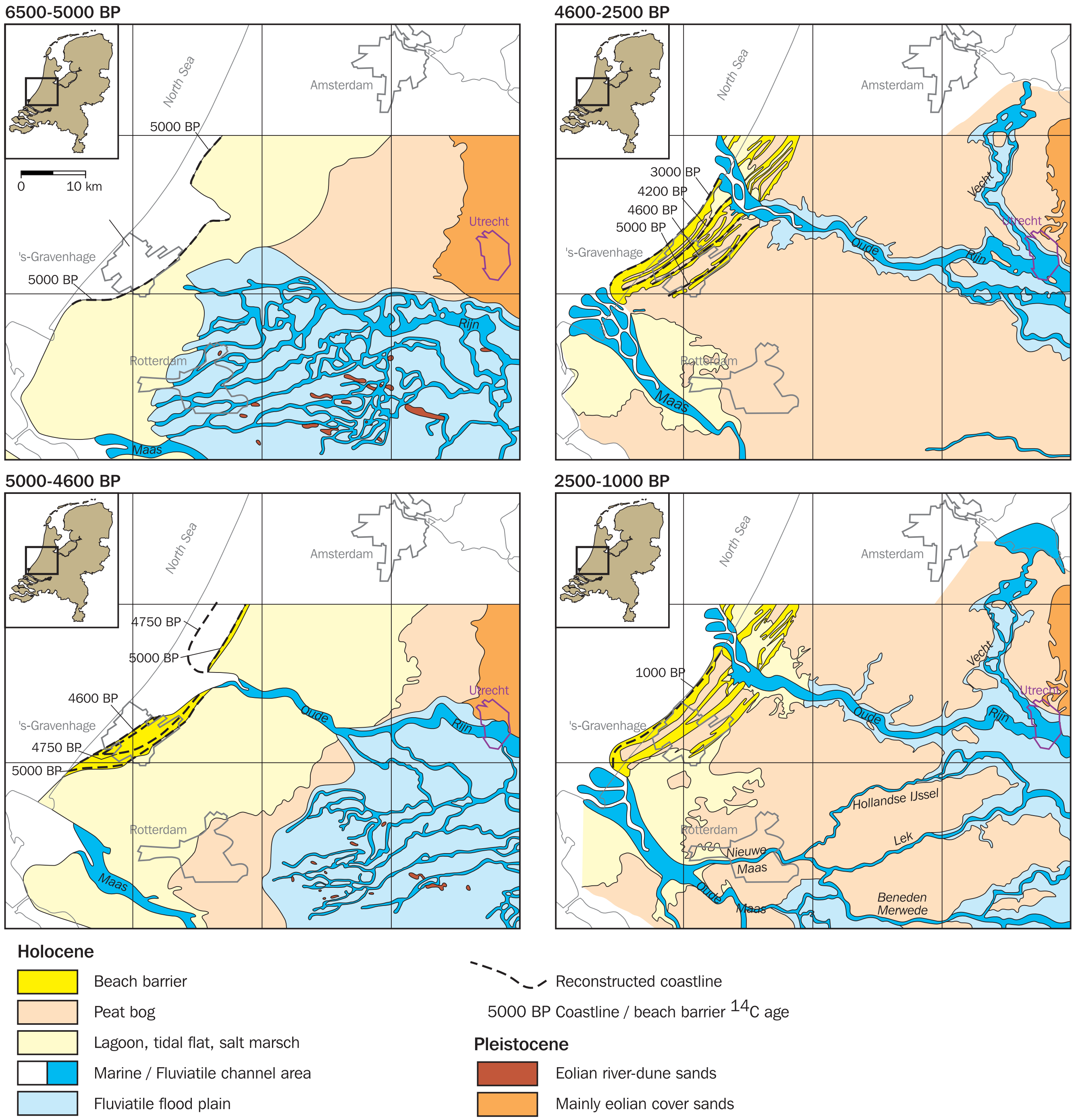

Fig. 14. Paleogeographic maps of the Middle and Late Holocene alluvial plain in the central-western Netherlands (after De Groot & De Gans, 1996).

ward, the influence of the sea-level rise on fluvial sedimentation decreased and the anastomosed river system gradually disintegrated. The Oude Rijn became the main Rhine channel, building a delta, and later an estuary, near Katwijk (Fig. 14b). The Oude Rijn remained active till about 1000 BP. The Meuse had its estuary near Hoek van Holland. The importance of the Rhine estuary near Katwijk decreased in favour of the Meuse estuary as other Rhine channels (Waal, Hollandse IJssel, Lek) found their way towards this estuary, depositing extensive clay beds on top of the peat. Other Rhine branches worked their way towards the north: the Utrechtse Vecht at about 4600 BP and the Gelderse IJssel at about 1600 BP, both depositing clays. The Vecht system, including the Angstel, comprises a sub-recent anastomosed system (Weerts et al., 2002). It was active until 1122 AD, when the Kromme Rijn was dammed near Wijk bij Duurstede, upstream of Utrecht.

Tributaries to the main Rhine and Meuse channels occur in brook valleys in the outcropping Pleistocene. In fact these brook valleys are misfit valleys. Their valley floors are remnants of Pleistocene braided channels into which meandering channels were cut that are comparable to the main river systems (De Gans, 1983; Cleveringa et al., 1988). During the Holocene the channels became filled with sand, clay and organic debris, while the former valley floors were covered by peat.

Inland peat

As discussed, the major peat beds were formed in the coastal areas since the late Atlantic period. In water-divide areas on the Pleistocene, inland peat started to accumulate earlier, from the Preboreal on, and extensive peat beds up to 10 m thick developed in areas such as south-east Groningen, parts of Drenthe and Overijssel, the Gelderse Vallei (Utrecht–Gelderland) and the Peel (Noord Brabant–Limburg; Fig. 13). The peat is generally composed of eutrophic peat at its base (wood peat – 'bosveen') and changes gradually into oligotrophic peat (*Sphagnum* peat – 'mosveen') towards the top (Zagwijn, 1986).

Man as a geological agent

Although early inhabitants already used local aggregates for their living, it is with the introduction of agriculture that major changes in the environment were introduced. From the beginning of the Subboreal, the area covered by forest slowly decreased as a result of spreading agriculture, and since the beginning of the Subatlantic, wide open spaces occurred and heath lands originated on a large scale. The first habitable mounds ('terpen') were built from about 1500 BP onward and the first dikes and canals date from the Roman period. The latter changed the erosion and sedimentation patterns of rivers and tidal channels considerably. The cultivation of peat lands led to a regional subsidence of about 5 m in the coastal zone, whereas agriculture on the Pleistocene areas led on the one hand to local soil erosion and on the other to more heath lands and to artificially raised soils ('essen').

The transition from an agricultural to an industrial society around 1850 accelerated the intensity of human interference with geological processes in such a way that after World War 2 the activity of man became a major geological factor.

Dikes

The construction of dikes along rivers decreased the deposition areas, with the result that the sedimentation in the land areas ('uiterwaarden') between the dikes increased their elevation substantially. On the other hand the adjacent areas behind the dikes subsided due to drainage. This, together with neglect of dike maintenance, led repeatedly to bursting of dikes and the washing out of numerous small lakes ('wielen'). The expected increases in sea level and in fluvial peak discharges will lead in the future to an increased flooding risk.

The construction of dikes along estuaries such as the Westerschelde in Zeeland impedes lateral erosion by the channels, but instead increases vertical erosion. This way, channels like the Westerschelde and the Marsdiep tidal inlet in Noord-Holland now attain depths of 65 and 50 m respectively. In contrast, the construction of sea dikes and breakwaters can also induce sedimentation. Examples are the silted-up areas at the seaside of the Oosterschelde Dam and the south side of the IJmuiden harbour breakwater west of Amsterdam.

The cultivation of peat lands

The drainage of peat lands gradually induced oxidation of the peat and ultimately led to subsidence. In the coastal zone the surface level of the peat lands decreased since about 1000 BP from about +4 m to –1 or –2 m NAP (Fig. 12). The subsidence still continues. In the coastal zone the excavation of peat resulted in lakes. Other lakes were formed as a result of marine erosion. Both kinds of lakes became enlarged by wind- induced wave erosion of their shores, giving rise to their SW-NE orientation.

From the 16th century on, part of these lakes were surrounded by dikes, pumped dry and converted into 'polders'. Such polders are 'artificial geological windows' where older tidal deposits are cropping out (Figs 12b, c). The land surface of these polders is generally about –4 to –5 m NAP. The deepest level is –6.72 m NAP in a polder north-east of Rotterdam, where man-induced subsidence since 1000 AD amounts to over 10 m.

The floodplains of the brook valleys, which were to a large degree composed of peat beds or other organic deposits, were also affected by subsidence, partly because of oxidation and decomposition of the peat, but also as a result of the extraction of iron oxides (bog iron ores; Van Montfrans et al., 1988).

Inland dunes

The introduction of agriculture since the beginning of the Subboreal, resulted in the increase of heath lands. Due to over-grazing, the vegetation cover was destroyed at places and the underlying Pleistocene sands, topped by a podzolic soil, were locally eroded. As a result, inland dunes were formed, which can be distinghuished from the Pleistocene eolian deposits by their darker, dirty colour.

Economic geology

Large quantities of surface minerals are or were extracted from the Quaternary in the Netherlands. They include peat, used as a fuel in the past (Van Bergen et al., this volume), and sand, gravel and clay which are applied in various ways in the construction of buildings, transport networks and dikes, and as far as sand is concerned also in sea-coast maintenance (Van der Meulen et al., this volume). More or less subordinate, historical resources are till, iron ore and diatomite, as well as salt obtained from peat. Furthermore, large quantities of fresh groundwater are obtained from aquifers in Quaternary sediments (De Vries, this volume).

Geo-conservation

From the late 19th and early 20th century on, the view of man upon his environment changed. In periods before, the natural landscape was mainly considered as an area to be cultivated. Gradually, uncultivated areas ('woeste gronden'), and especially their flora and fauna, became regarded as 'nature', and 'nature areas' needed to be preserved, protected or restored. Much later, localities of earth-scientific interest also began to receive attention. In 1989, Gonggrijp presented a paper listing 119 of such geo-sites. Most of these represent geomorphological phenomena. De Gans (2000) indicated the importance of sediments in specific sites as earth-scientific archives.

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Magmatism in the Netherlands: expression of the north-west European rifting history

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ABSTRACT

Wells drilled during the exploration for hydrocarbons have revealed the significance of magmatic activity in the Dutch geological history. Igneous rocks were encountered in more than 60 wells, mainly concentrated in the eastern provinces of the country, the western onshore and offshore area, and the northern offshore. Intrusive rocks, all but one emplaced in Carboniferous and younger sediments, dominate over extrusive rocks. A prominent exception is the Jurassic Zuidwal Volcano under the Waddenzee, where thick extrusive deposits and associated subsurface features define a well-developed volcanic centre. Radiometric age dating indicates that the timing of magmatism in the Netherlands largely coincides with the Late Paleozoic and Mesozoic patterns of magmatic activity that affected the North Sea region and adjacent areas. Analysed samples suggest that virtually all igneous rocks encountered have a mafic composition and moderate to high alkali contents. Geochemical signatures are consistent with a within-plate tectonic setting throughout the successive episodes of magmatic activity, amongst which the Early Permian and Mid Jurassic to Early Cretaceous phases were the main ones. The lithospheric rifting processes that have dominated the north-west European geological history since the Paleozoic provided favourable conditions for melt generation, the emplacement of intrusive magma bodies and associated volcanism.

Keywords: North Sea Basin, intrusive rocks, volcanism, igneous geochemistry, within-plate magmatism

Introduction

The flat landscape and sediment-dominated deltaic environment of the Netherlands are not easily associated with the presence of igneous rocks and magmatic processes. Nevertheless, the country's subsurface hosts a surprising record of magmatic events that cover large parts of its geological history. The first igneous rocks were discovered in 1923 in the Corle exploration well, drilled in the eastern part of the province of Gelderland, where a dolerite was encountered (Tesch, 1925, 1928; Tomkeieff & Tesch, 1931). Several years later, three similar intrusions were found in the nearby Meddeho and Hupsel wells (Tesch & Van Voorthuysen, 1944). All these intrusions occur in Carboniferous shales at depths between 957 and 1320 m. Since then, numerous hydrocarbon exploration wells have encountered intrusive or extrusive igneous rocks, both onshore and offshore. The most pronounced igneous feature is the Late Jurassic Zuidwal complex under the Waddenzee, which constitutes the only well-defined volcanic centre in the Dutch subsurface.

Recently released data from onshore and offshore wells have permitted to compile a comprehensive inventory of igneous rocks in the Dutch subsurface, including details on their stratigraphic framework and new K-Ar ages (Sissingh, 1986, 2004). Collectively, the occurrences appear to fit into the generalised pattern of magmatism that has been recognised in other parts of north-west Europe, and that can be linked to the large-scale tectonic evolution in this region (cf. Woodhall & Knox, 1979; Dixon et al., 1981; Ziegler, 1990, 1992; Latin & Waters, 1992). Here we present an overview of currently available information on the distribution, chronology and compositional characteristics, with particular emphasis on magmagenetic aspects that are relevant in the context of the Late Paleozoic-Mesozoic rift history of north-west Europe.

Throughout this study, reference is made to structural elements that are commonly shown on geological maps of the Dutch subsurface. We use their names in connection with the igneous occurrences as a geographic reference only, without implying genetic links. An overview of the structural history of the Netherlands can be found in De Jager (this volume). Rock names mentioned are not based on a rigorous application of petrographic classifications, but are generally those used in the original publications. Also, qualifications concerning an intrusive or extrusive mode of emplacement of the igneous bodies should be considered with care, as interpretations are sometimes based on poor or conflicting evidence.

Igneous rocks and volcanogenic sediments in the Netherlands and adjacent areas

The distribution and structural setting of about 60 wells in which igneous rocks have been identified in the Netherlands are shown in Figures 1 and 2. Samples were obtained

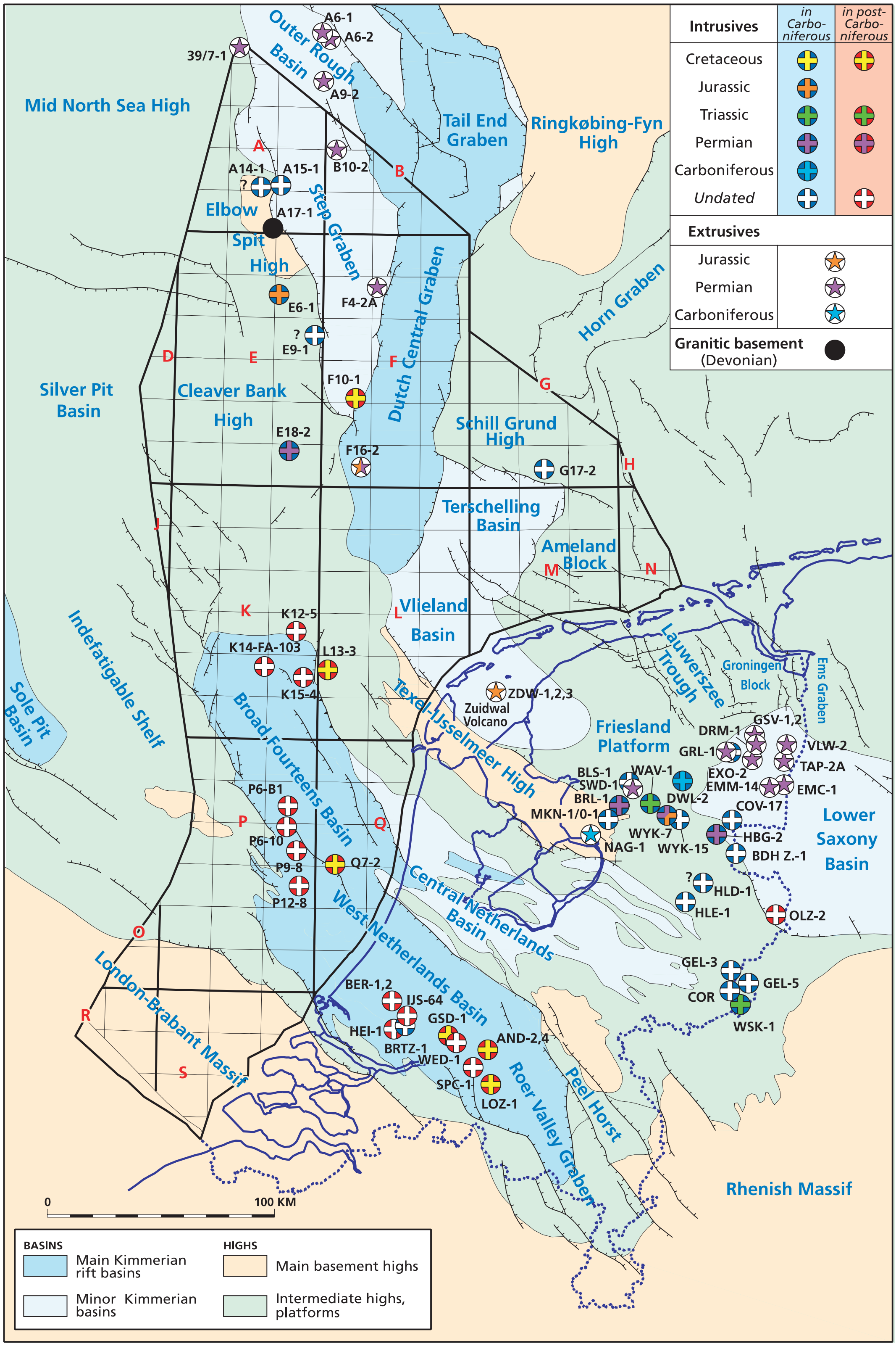

Fig. 1. Structural map (De Jager, this volume) showing locations of wells with igneous rocks in the Netherlands; these are listed in Table 1. Well A17-1 (black circle) bottomed in dated granitic basement and also penetrated dated volcanics in Devonian sediments.

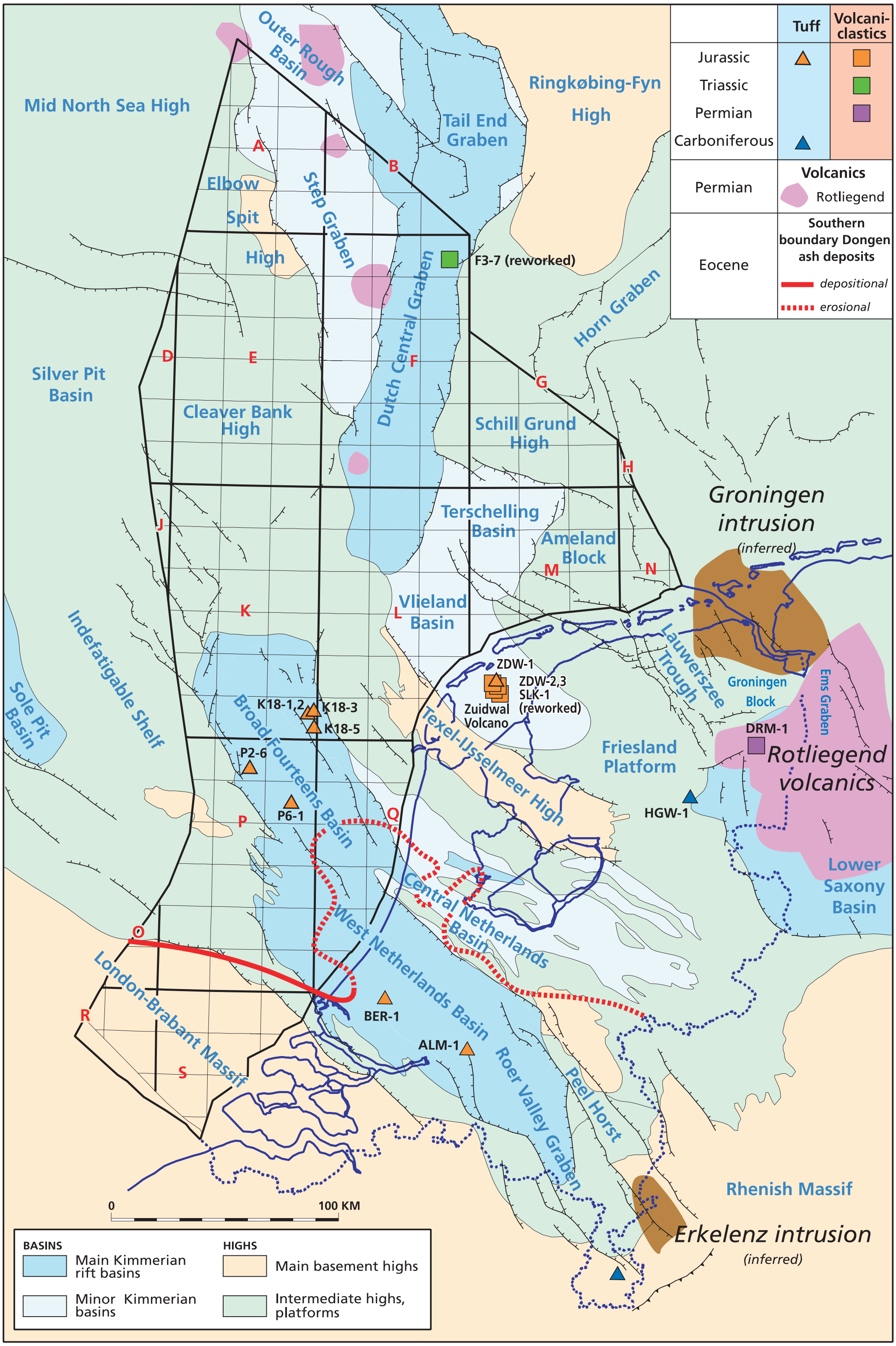

Fig. 2. Structural map (De Jager, this volume) showing locations of wells with tuffs and volcaniclastics (Table 1), distribution of Rotliegend volcanics and inferred igneous intrusions, and southern boundary of Dongen ash deposits.

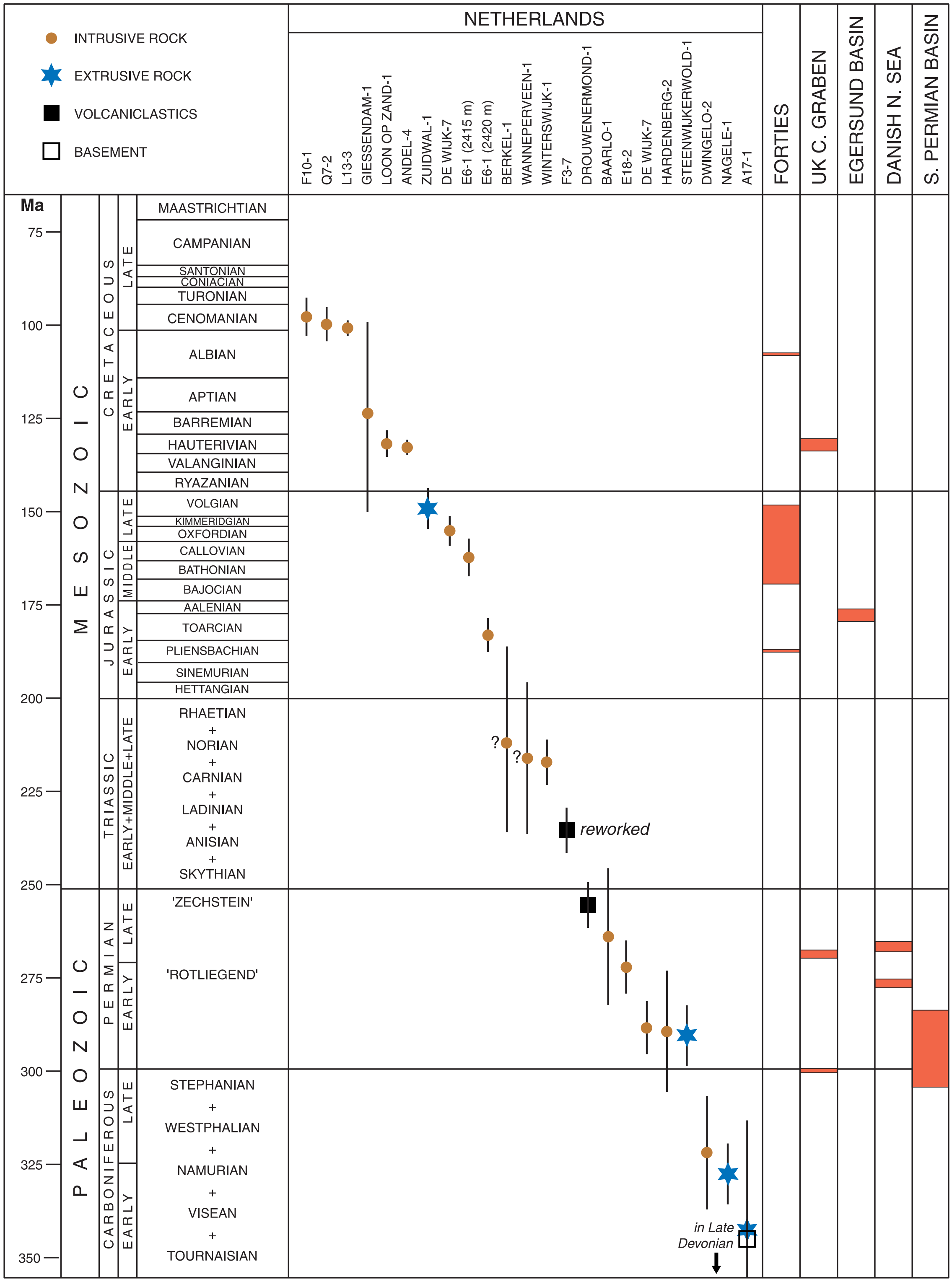

Fig. 3. Stratigraphic overview of younger Paleozoic and Mesozoic, radiometrically dated igneous and volcaniclastic rocks in the Netherlands and adjacent NW European areas (see Table 1 for references).

from depths between 1200 and 4500 m, and about 30% were dated radiometrically (Table 1, Fig. 3). Most of the dated and undated rocks are intrusives. Because original petrographic descriptions are sometimes lacking or inconclusive, some rocks were only classified as lava if their ra-

Table 1a. Radiometrically dated igneous rocks in the Netherlands subsurface.

Age, location	Well	Code	Dated interval (m)	Age (Ma)	Method	Stratigr. interval	Emplace-ment	Intruded interval	Rock type/ composition	Refs.	Remarks
Cretaceous											
Rim Dutch Central Graben	*Offshore*	F10-1 (=PL1)	3426-3460 T.D.	99 ± 5	K-Ar	'Mid-' Cretaceous	I	Zechstein	Lamprophyre	1, 2	
Rim Broad Fourteens Basin	*Offshore*	L13-3	~1820	101 ± 1	$^{40}Ar/^{39}Ar$	'Mid-' Cretaceous	I	Zechstein	Olivine nephelinite	3	(a)
Broad Fourteens Basin	*Offshore*	Q7-2	~3450(?)	95 ± 2 to 106 ± 2	$^{40}Ar/^{39}Ar$	'Mid-' Cretaceous	I	Triassic	Undersaturated basalt	3	
West Netherlands Basin	Giessendam-1	GSD-1	1190-1218	125 ± 25	K-Ar	Early Cretaceous	I	Jur-Cret transition	'Basalt'	1	
West Netherlands Basin	Loon-op-Zand-1	LOZ-1	2572-2610 (prob)	132 ± 3	$^{40}Ar/^{39}Ar$	Early Cretaceous	I	Lower Jurassic	Nephelinite, basanite	3	
West Netherlands Basin	Andel-4	AND-4	1651.4-1652.1 or 1657.1-1658.0	133 ± 2	$^{40}Ar/^{39}Ar$	Early Cretaceous	I	Middle Jurassic	Nephelinite, basanite	3	
Jurassic											
Vlieland Basin	Zuidwal-1	ZDW-1	1944-3002 T.D.	145	K-Ar	Late Jurassic	E		Trachyte	4	(b)
				144 ± 1	$^{40}Ar/^{39}Ar$				Phonolite, biotite pyroxenite,	3	(c)
				152 ± 3	$^{40}Ar/^{36}Ar$-$^{40}K/^{36}Ar$				Trachyte, phonolite, leucitite	5	
East of Texel-IJsselmeer High	De Wijk-7	WYK-7	2443-2486	155 ± 4	K-Ar	Late Jurassic	I	Silesian	Olivine gabbro	1, 6	
Rim Step Graben	*Offshore*	E6-1 (2415 m)	2405-2423	161 ± 4	K-Ar	Middle Jurassic	I	Dinantian	*No data*	1	
		E6-1 (2420 m)		183 ± 4	K-Ar	Early Jurassic	I	Dinantian	*No data*	1	
Triassic											
West Netherlands Basin	Berkel-1	BER-1	~2944-2970	214 ± 25?	K-Ar	Late Triassic	I	Lower Jurassic	Essexite or theralite	1, 3, 7	(d)
East of Texel-IJsselmeer High	Wanneperveen-1	WAV-1	2030.5-2035	217 ± 20?	K-Ar	Late Triassic	I	Silesian	Dolerite, (olivine-) gabbro	1, 8	(e)
Achterhoek High	Winterswijk-1	WSK-1	4089.5-4149.5	218 ± 6	K-Ar	Late Triassic	I	Silesian	Olivine dolerite	1, 9	
Dutch Central Graben	*Offshore*	F3-7	2928.0-2928.1	236 ± 6	K-Ar	Middle Triassic	V		*No data*	1	
Permian											
Lower Saxony Basin	Drouwenermond-1	DRM-1	3920.5-3953.5	258 ± 6	K-Ar	Late Permian	V		*No data*	1	
East of Texel-IJsselmeer High	Baarlo-1	BRL-1	1772.5-1775	266 ± 18	K-Ar	Early Permian	I	Silesian	Dolerite ('diabase')	1	
Cleaver Bank High	*Offshore*	E18-2	4437-4452	274 ± 7	K-Ar	Early Permian	I	Dinantian	Gabbro	1	
East of Texel-IJsselmeer High	De Wijk-7	WYK-7	2684-2691	289 ± 7	K-Ar	Early Permian	I	Silesian	Gabbro	1	
Rim Lower Saxony Basin	Hardenberg-2	HBG-2	3376.5-3441	290 ± 16	K-Ar	Early Permian	I	Silesian	Olivine gabbro	1	
East of Texel-IJsselmeer High	Steenwijkerwold-1	SWD-1	1937.5-1944.5	291 ± 8	K-Ar	Early Permian	E		'Basalt'	1	(f)
Carboniferous											
East of Texel-IJsselmeer High	Dwingelo-2	DWL-2	3754.3-3792.2 T.D.	322 ± 15	K-Ar	Carboniferous	I	Silesian	Olivine gabbro	1, 6	
Eastern Texel-IJsselmeer High	Nagele-1	NAG-1	2772.5-2776	327 ± 8	K-Ar	Namurian	E		*No data*	1	(g)
Devonian											
Elbow Spit High	*Offshore*	A17-1	2157-2195	341 ± 30	?	Late Devonian	E		Rhyolite-Rhyodacite	1	(h)
			3013-3043.7 T.D.	Min. 346 ± 7	$^{40}Ar/^{39}Ar$	'Basement'	I		Biotite monzo-granite	10	(I)

[1] For notes see Table 1b.

Table 1b. Undated igneous rocks in the Netherlands subsurface.

Location	Well	Code	Stratigr. interval	Emplace-ment	Intruded interval	Rock type/ composition	Refs.	Remarks
Achterhoek High	Corle	COR		I	Silesian	Dolerite	11	
	Gelria-3 (Hupsel)	GEL-3		I	Silesian	Dolerite ('diabase')	12	
	Gelria-5 (Meddeho)	GEL-5		I	Silesian	Dolerite ('diabase')	12	(j)
Lower Saxony Basin	Balderhaar Z.-1	BDH Z.-1		I	?	*No data*	1	
	Coevorden-17	COV-17		I	Silesian	*No data*	1	
	Emmen-14	EMM-14	Early Permian	E		*No data*	1	
	Emmer-Compascuum-1	EMC-1	Early Permian	E		Basalt and tephra	1, 13	(k)
	Exloo-2	EXO-2	Early Permian	E		*No data*	1	
	Gasselternijveen-1	GSV-1	Early Permian	E		*No data*	1	
	Gasselternijveen-2	GSV-2	Early Permian	E		*No data*	1	
	Grollo-1	GRL-1		I	Silesian	Dolerite ('diabase')	1	
	Oldenzaal-2	OLZ-2		I	'Wealden'	Hornblende diabase	14	
	Ter Apel-2A	TAP-2A	Early Permian	E		*No data*	1	
	Vlagtwedde-2	VLW-2	Early Permian	E		*No data*	1	
W of L. Saxony Basin	Haarle-1	HLE-1		I	Silesian	Dolerite ('diabase')	1	
	Hellendoorn-1	HLD-1		?	?	*No data*	1	
	Hoogenweg-1	HGW-1	Silesian	Ash layer		*No data*	15	(l)
E of Texel-IJsselm. High	De Blesse-1	BLS-1		I	Silesian	*No data*	1	
	De Wijk-15	WYK-15		I	Silesian	*No data*	1	
	Marknesse 1/O-1	MKN-1/O-1		I	Silesian	*No data*	1	
West Netherlands Basin	Almkerk-1	ALM-1	Jur-Cret transition	Tuff		*No data*	1	
	Andel-2	AND-2		I	Middle Jurassic	Olivine nephelinite, basanite	2	
	Barendrecht-Ziedewij-1	BRTZ-1		I	Silesian	Dolerite	15	
	Berkel-1	BER-1	Jur-Cret transition	Tuff		*No data*	1	
	Berkel-2	BER-2		I	?	*No data*	1	
	Heinenoord-1	HEI-1		I	U. Jurassic-L. Cret	Alkali basalt	16	
	Sprang-Capelle-1	SPC-1		I(?)	Triassic	Contains nepheline and biotite	17	(m)
	Werkendam-1	WED-1		I	Middle Jurassic	Dolerite	1	
	IJsselmonde-64	IJS-64		I	Lower Jurassic	Basanite	1	
Vlieland Basin	Zuidwal-2	ZDW-2	Portlandian-Ryazanian	V		*No data*	18	(n)
	Zuidwal-3	ZDW-3	Portlandian-Ryazanian	V		*No data*	18	(n)
	Slenk-1	SLK-1	Portlandian-Ryazanian	V		*No data*	18	(n)

(Continued.)

Table 1b. Continued.

Location	Well	Code	Stratigr. interval	Emplace-ment	Intruded interval	Rock type/ composition	Refs.	Remarks
Step Graben	*Offshore*	A14-1		I	Silesian	Porphyry ?	1	
	Offshore	A15-1		I	Silesian	Gabbro, micro-granodiorite	2	(o)
	Offshore	B10-2	Early Permian	E		No data	1	
	Offshore	E9-1		I (?)	Silesian	Porphyry?	1	
Rim Central Graben	*Offshore*	F4-2A		E ?	Rotliegend	Micro-granodiorite	2	
	Offshore	F16-2	Jur-Cret transition	E		*No data*	19	(p)
			Late Permian	E		Rhyolitic ?	19	(q)
Schill Grund High	*Offshore*	G17-2	Permian ?	I	Silesian	Dolerite	2	
Broad Fourteens Basin	*Offshore*	K12-5		I	Zechstein	Lamprophyre	2	
	Offshore	K14-FA-103		I(?)	Lower Permian	Trachybasalt	3	
	Offshore	K15-4		I(?)	?	*No data*	1	
	Offshore	K18-1, 2	Jur-Cret transition	Tuff		*No data*	1	
	Offshore	K18-3	Jur-Cret transition	Tuff		*No data*	1	
	Offshore	K18-5	Jur-Cret transition	Tuff		*No data*	1	
	Offshore	P2-6	Jur-Cret transition	Tuff		*No data*	1	
	Offshore	P6-1	Jur-Cret transition	Tuff		*No data*	1	
	Offshore	P6-B1		I	Triassic	Lamprophyre	2	
	Offshore	P6-10		I(?)	Triassic	Mafic rock	1	(r)
	Offshore	P9-8		I(?)	Triassic	Olivine basalt	1	(s)
	Offshore	P12-8		I(?)	Triassic	Lamprophyric	1	(t)

[1]For locations see Figure 1. T.D.: terminal depth of well. Emplacement: E: extrusive, I: intrusive, V: volcaniclastics. Note that qualifications concerning an extrusive (lava) or intrusive (e.g. sill, dyke) character should be considered with care, as available information is sometimes insufficient or inconclusive. Likewise, the quality of the radiometric ages is variable and sometimes difficult to evaluate in the absence of sufficient documentation. The possibility of a significant error should be taken into account, particularly for K-Ar ages obtained on bulk rocks (ref. 1) that have suffered some degree of alteration, as is the case in many of these rocks. References: 1: Sissingh (2004), 2: Kuijper (1991), 3: Dixon et al. (1981), 4: Harrison et al. (1979), 5: Perrot & Van der Poel (1987), 6: Eigenfeld & Eigenfeld (1986), 7: Van der Sijp (1953), 8: Kimpe (1953), 9: NITG (1998), 10: Frost et al. (1981), 11: Tomkeieff & Tesch (1931), 12: Tesch & Van Voorthuysen (1944), 13: NITG (2000), 14: Van Voorthuysen (1944), 15: Van Buggenum & Den Hartog Jager (this volume), 16: Helmers (1991), 17: Personal communication Wintershall, 18: Herngreen et al. (1991), 19: AMOCO. Remarks: (a) Overprint at 51.9 $\pm$ 0.3 Ma; (b) See text for younger ages; (c) Also phonolitic basanite, leucite basanite, leucite tephrite; (d) K-Ar age must be incorrect; fine-grained; ref. 7: hornblende basalt; (e) Questionable age, see text; (f) K-Ar age not consistent with extrusive nature; (g) RGD (1993) reports swarm of thin dolerite-like intrusive bodies; (h) Presumably present as stack of flows; (i) 'Minimum' age = alteration?; (j) Two intrusions; (k) ~75 m, alternation of basalt and tephra layers; (l) 5-cm-thick ash layer; (m) Two levels in Solling and Sleen shale. May be extrusive or reworked; (n) Probably reworked; (o) Bimodal magmatism; (p) Tentatively correlated with Zuidwal Volcanic Formation; (q) Underlain and covered by Zechstein evaporites; (r) Altered volcanic breccia?; (s) Highly altered; (t) Based on loose crystals, mainly cpx, from hand-picked cuttings.

diometric dates corresponded with the age of the sedimentary interval in which they occur (e.g. A17-1, Nagele-1). Intrusive bodies occurring in Carboniferous sediments are concentrated in the east of the Netherlands and in the northern offshore. Those emplaced in post-Carboniferous strata are largely restricted to the West Nederlands and Broad Fourteens basins. Wells in the same region contain most of the recorded tuffaceous sediments. Such deposits also occur in Hoogenweg-1, west of the Lower Saxony Basin, and Zuidwal-1 in the Vlieland Basin. In the latter well, more coarse-grained volcaniclastics have been found as well. Reworked volcaniclastics occur in the F3-7 well in the Dutch Central Graben. Below follows a description of the Dutch igneous rocks and volcanogenic deposits with the available age constraints in stratigraphic order, together with brief accounts on their possible equivalents in north-west Europe.

Crystalline basement

The oldest igneous rocks in the Netherlands belong to the crystalline basement, which has been encountered only in offshore well *A17-1* on the Elbow Spit High (Fig. 1). According to Frost et al. (1981), the rocks can be classified as a biotite monzo-granite containing heavily altered biotite and oligoclase. These authors reported a $^{40}Ar/^{39}Ar$ age of 346 ± 7 Ma, obtained on micas, which should be regarded as a minimum in view of the degree of alteration. This granite, which is covered by Devonian sediments, was presumably emplaced during or shortly after the Caledonian orogeny.

The *A17-1* granite can be assigned to the Caledonian crystalline basement complex that has been revealed by about 30 wells in the North Sea Basin (Frost et al., 1981). These occurrences represent a belt of medium to high-grade metamorphic and intrusive rocks, which links the Caledonides of Scotland and Norway. Moreover, data from boreholes in Denmark and north Germany point to a connection between the North Sea and the Polish Caledonides (Ziegler, 1990; Pharaoh, 1999).

Available $^{40}Ar/^{39}Ar$ dates of these basement rocks cluster around 440 to 450 Ma (earliest Silurian or latest Ordovician) and fall within the Caledonian radiometric dates of Britain and Scandinavia (Frost et al., 1981, and references therein). Many of the ages obtained are likely to represent overprints of earlier (Grampian or older, Precambrian) phases of metamorphism or deformation. Post-Caledonian overprints show less consistency and can be attributed to local thermal or tectonic events. One such event around 350 Ma may explain the apparent age of the A17-1 monzo-granite.

Devonian

The *A17-1* well is also the site of Devonian igneous rocks consisting of altered rhyolitic volcanics or quartz-porphyry. These rocks are intercalated in non-metamorphic Old Red Sandstone, which covers the basement, and have been dated at 341 ± 30 Ma (see Sissingh, 2004). If the extrusive mode of emplacement and stratigraphic position are correct, this Early Carboniferous radiometric age is too young. In the North Sea, this is a rather isolated example of the volcanic activity that accompanied the tensional tectonics, which characterised the Devonian development of north-west and central Europe. Devonian bimodal alkaline volcanics are present in the Cornwall and Rhenish basins, and also occur in the more southerly Central Armorican – Saxothuringian Basin. In contrast, calcalkaline volcanism and emplacement of post-orogenic granites accompanied the development of Old Red basins in the northern British Isles (Ziegler, 1990).

Carboniferous

Igneous rocks with a radiometrically determined Carboniferous age are restricted to onshore wells around the eastern Texel-IJsselmeer High. A basaltic lava from *Nagele-1* has been dated at 327 ± 8 Ma, which is consistent with its occurrence in the Namurian-Westphalian sequence (Sissingh, 2004). An extrusive basaltic rock in Westphalian sediments in *Steenwijkerwold-1*, however, was dated as Early Permian, which conflicts with its extrusive nature (Sissingh, 2004; Van Buggenum & Den Hartog Jager, this volume). Gabbroic intrusive rocks in *Dwingelo-2* (Thiadens, 1963; Eigenfeld & Eigenfeld, 1986) form a sill-like body of several kilometres length according to seismic data. K-Ar dating yielded an age of 322 ± 15 Ma. Widespread kaolinite-coal-tonstein beds in the Limburg Group represent altered tuffaceous horizons in the former coal-mining area of the southern Netherlands, providing evidence for explosive volcanism during the Late Carboniferous (Lippolt et al., 1984). A 5-cm-thick ash layer at 3134 m in *Hoogenweg-1* (Van Buggenum & Den Hartog Jager, this volume) may be a northern equivalent. Early Carboniferous volcanism is evident in the Cornwall Basin–Rhenish Basin and the graben systems of the British Isles (see Timmerman, 2004), whereas intense Late Carboniferous calcalkaline syn-orogenic volcanism characterised the internal Variscides of western and central Europe, in which post-orogenic magmatic activity persisted during the Stephanian and Early Permian (Ziegler, 1981, 1990; Ziegler et al., 2004, and references therein).

Permian

Placing the Carboniferous-Permian boundary at 299 Ma (Gradstein et al., 2004), igneous rocks of Early Permian age occur around the eastern part of the Texel-IJsselmeer High, in the Lower Saxony Basin and in the Dutch Central Graben area (Figs 1, 2; see also Geluk, this volume). Rocks that constitute the Emmen Volcanic Formation onshore consist of basaltic lava flows and pyroclastics reach-

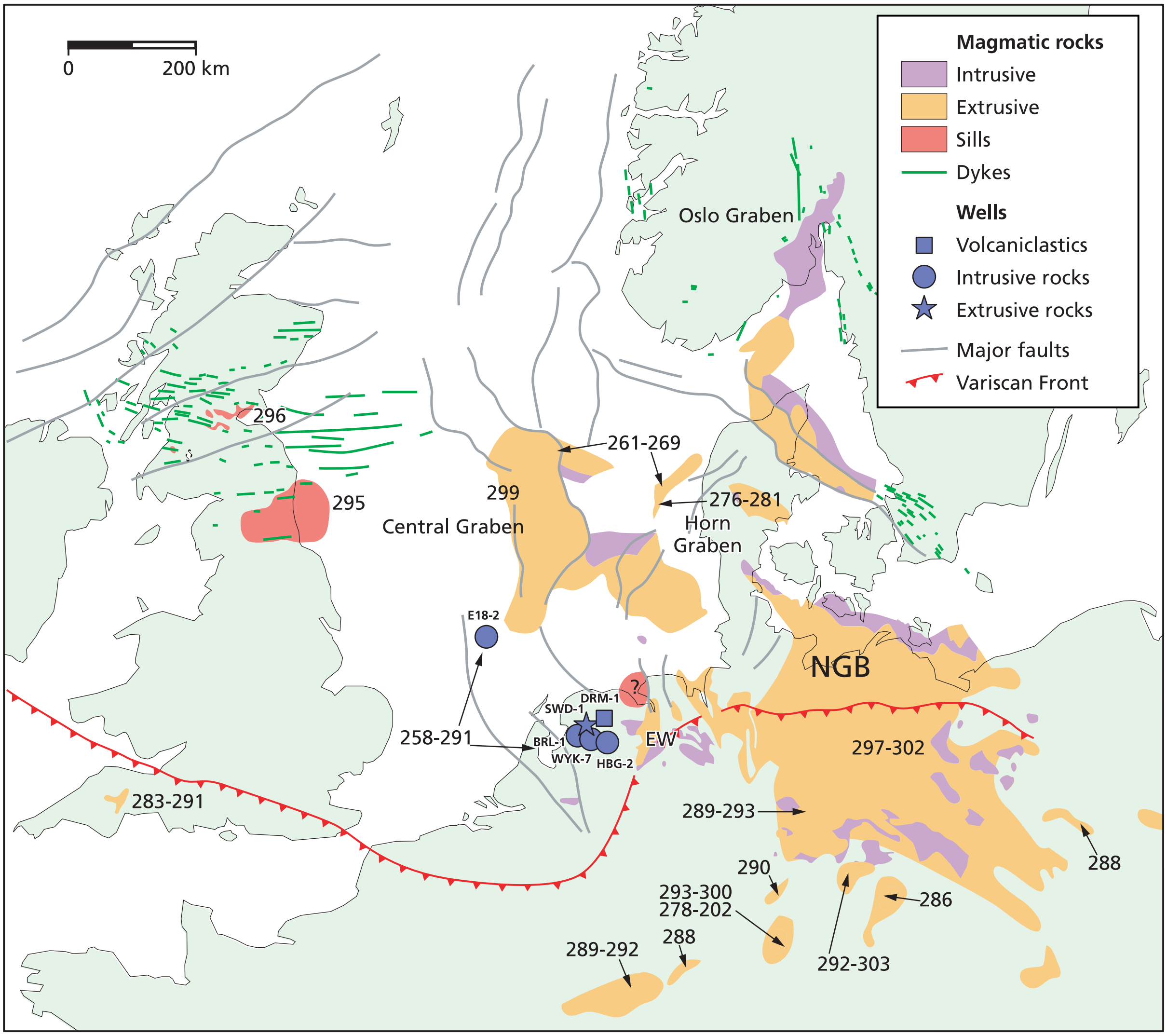

Fig. 4. Distribution of Permian and latest Carboniferous volcanic rocks in NW Europe. Numbers refer to radiometric ages in Ma (Sissingh, 1986, 2004; Glennie, 1997, 1998, and references therein; Breitkreuz & Kennedy, 1999; Stemmerik et al., 2000; Heeremans et al., 2004). EW: Ems-Weser area, NGB: North-east German Basin.

ing a maximum thickness of 80 m (NITG, 2000). Mafic intrusive rocks in *Hardenberg-2* and *De Wijk-7* yielded K-Ar ages of ca. 290 Ma (Sissingh, 2004; Table 1), falling within the age range of the Lower Rotliegend volcanics in north Germany that are confined to the Early Permian rather than to the Stephanian (Lippolt & Hess, 1989; Plein, 1995; Glennie, 1997; cf. Fig. 4). Somewhat younger dated occurrences are a basaltic intrusive in *Baarlo-1* (266 ± 18 Ma) and a volcaniclastic deposit in *Drouwenermond-1* (258 ± 6 Ma).

Offshore occurrences of Permian magmatic rocks are located in the Cleaver Bank High (*E18-2* well; gabbro dated at 274 ± 7 Ma) and near the Dutch Central Graben (e.g. a porphyritic rhyolite in well *F4-2A* and rocks in wells in and near blocks A and B; Figs 1, 2). The Central Graben volcanics are nearly 150 m thick and consist of pyroclastics and of lavas up to tens of metres thick (Geluk, 1997). Kuijper (1991) described a dolerite intrusion in the Upper Carboniferous of well *G17-2*. Petrological data suggest an affinity with Permian rather than with Jurassic-Cretaceous magmatism, but radiometric data are lacking.

On the basis of petrographic similarity, Dixon et al. (1981) proposed that most, if not all of 15 basaltic bodies in *Wanneperveen-1*, referred to as intrusives by Kimpe (1953), are Permian flows equivalent to those in the North Sea wells. The K-Ar dating of a Wanneperveen-1 sample (interval 2030.5-2035 m), however, yielded an age of 217 ± 20 Ma (Sissingh, 2004), making it clearly younger than the Rotliegend volcanics.

West of Denmark, Dixon et al. (1981) reported the presence of a bimodal association of Lower Permian volcanics along the northern and western flanks of the *Ringkøbing*

High and from within the *Horn Graben*. Strongly altered basalts contain primary biotite and have been classified as transitional between olivine tholeiites and mildly alkaline basalts on the basis of immobile trace elements. The silicic rocks may represent silicified trachytes, although their original chemical composition is not clear. More recently, Stemmerik et al. (2000) suggested that Rotliegend volcanism in the Danish part of the North Sea took place during two separate events: 276-281 Ma and 261-269 Ma, post-dating the Lower Rotliegend volcanism in the Southern Permian Basin (Fig. 4). However, Heeremans et al. (2004) argue that these K-Ar ages might be too young, based on a new Ar-Ar age of 299 ± 3 Ma they obtained on a basalt sample from well 39/2-4 on the adjacent western flank of the UK Central Graben. Different periods of extensive intrusive and extrusive magmatism accompanied the development of the *Oslo Rift* (Neumann et al., 1992, 2004). The igneous activity, which started at ca. 305 Ma and continued into the Triassic, has a strongly alkaline character varying between basaltic and granitic. A comprehensive study of the widespread Late Carboniferous-Permian magmatism in the *North-east German Basin* showed that thick rhyolitic rocks and ignimbrites clearly dominate in volume over intermediate and basaltic varieties in this part of the Southern Permian Basin (Benek et al., 1996). Zircon dating yielded ages of 294 to 307 Ma for the volcanic activity in this area, thus straddling the Carboniferous-Permian boundary (Breitkreuz & Kennedy, 1999). Interestingly, the authors also report populations of 'old' zircons that include ages reflecting reworked Cadomian and older Gondwanan elements.

The Early Permian igneous rocks in the Netherlands thus represent the same magmatic activity that produced the Lower Rotliegend volcanics in other parts of the Southern Permian Basin, which extends over some 1700 km from England across the North Sea through northern Germany into Poland (cf. Ziegler, 1990). Figure 4 shows the distribution of the Permian volcanics, the thickest of which may have been associated with caldera subsidence (Benek et al., 1996; Glennie, 1997, and references therein; cf. Scheck & Bayer, 1999). The occurrences in the eastern part of the Texel-IJsselmeer High and nearby areas, including the undated rocks of the Emmen Formation, can be considered as a westerly extension of the Rotliegend volcanics in the adjacent German Ems Low ('Ems Graben' in Figs 1, 2). These have been described as spilitised, mildly alkaline andesites and subordinate basaltic and rhyolitic rocks (Eckhardt, 1979; cf. Plein, 1995, and references therein). Eigenfeld & Eigenfeld (1986) used petrographic criteria to infer that the 'permo-carboniferous' mafic intrusive rocks in the eastern Netherlands represent the continuation of the Ems Low volcanics. However, according to currently available age data, only the 'olivine gabbro' of De Wijk-7 (2684-2691 m) would be coeval, whereas the other 'olivine gabbros' are older (Dwingelo-2) or younger (Wanneperveen-1, provided its K-Ar age of 217 ± 20 Ma is correct). The Corle 'melaphyre' and Gelria-3 (Hupsel) 'leucophyre' intrusives, originally referred to as 'dolerites' (Tomkeieff & Tesch, 1931; Tesch & Van Voorthuysen, 1944), and the Gelria-5 (Meddeho) 'dolerite' rocks have not been dated.

Triassic

Only minor magmatic activity accompanied basin development in north-west Europe during most of the Late Permian and Triassic. Traces of Zechstein volcanism have been found in the eastern parts of the Mid North Sea High (Ziegler, 1990). The main episode of dyke intrusions in the *Sunnhørdland* region in west Norway is expressed by alkaline, probably basic rocks of Triassic age (229 to 208 Ma; Faerseth et al., 1976). These probably extend into the Middle Jurassic. Dykes of Triassic age (219 ± 6 Ma) also occur in the *Oslo Graben*. They represent the end of the igneous activity since Late Carboniferous times (Neumann et al., 1992). Late Triassic-Early Jurassic dykes are also present in *north-west Brittany* (France), but these rocks probably have a low-alkali tholeiitic affinity (references in Harrison et al., 1979).

Three rocks from Dutch onshore wells yielded Triassic ages of 214 to 218 Ma according to K-Ar dating (Sissingh, 2004). Olivine basalt was found between the Posidonia and Werkendam shales in *Berkel-1* in the West Netherlands Basin. As these shales are Toarcian and Aalenian-Bajocian in age, the basalt must post-date the Early Jurassic, which discredits its Triassic radiometric date. Van der Sijp (1953) reported a 48-m-thick, altered hornblende basalt in this well, referring to it as an intrusive rock associated with a quartz-porphyritic rock, probably a dyke. Petrophysical data indicate that these rocks occur in two distinct intervals. In the absence of information on the depth of his samples, it cannot be verified whether the petrographic description of Van der Sijp (1953) corresponds to the dated rock. Dixon et al. (1981) suggested that this hornblende basalt may be younger (see below). The dated sample in *Wanneperveen-1* forms part of a series of metasomatically altered doleritic and gabbroic intrusives described by Kimpe (1953). The author inferred that these basalts and dolerites occur as dykes, whereas the gabbros may constitute a relatively large laccolithic body. The mode of emplacement of a rock in *Winterswijk-1*, dated at 218 ± 6 Ma (Sissingh, 2004), is unknown, but its presence in the older Namurian-Westphalian sequence points to an intrusive nature. Undated, hydrothermally altered doleritic dykes have been described in three nearby wells in eastern Gelderland, viz. *Corle, Gelria-3 (Hupsel)* and *Gelria-5 (Meddeho)*. All these dykes intruded Carboniferous shales (Tomkeieff & Tesch, 1931; Tesch & Van Voorthuysen, 1944). A Triassic age for these rocks is conceivable but would not be compatible with the hypothesis that they form part of the Early Permian Ems Low volcanic province, as proposed by Eigenfeld & Eigenfeld (1986).

The only offshore well with a Triassic igneous deposit is *F3-7* in the Dutch Central Graben. It contains volcaniclastics that yielded a K-Ar age of 236 ± 6 Ma (Sissingh, 2004). However, undated basaltic and lamprophyric rocks, encountered in wells along the Broad-Fourteens Basin (*P6-B1, P6-10, P9-8, P12-8*; Fig. 1), occur in Triassic stratigraphic intervals. Because their supposedly intrusive nature is insufficiently confirmed by available data, an extrusive origin cannot be excluded, which may imply that Triassic magmatism was more widespread. A similar uncertainty holds for nepheline and biotite-bearing rocks in the *Sprang-Capelle-1* well in the West Netherlands Basin that are also intercalated in Triassic sediments.

Jurassic

THE ZUIDWAL VOLCANIC CENTRE

The most prominent volcanic feature in the Netherlands is the 'Zuidwal Volcano', identified during exploration below the Waddenzee (Cottençon et al., 1975). This volcanic centre is located north of the Texel-IJsselmeer High in the Vlieland Basin (Fig. 1). Cores from the *Zuidwal-1* well revealed volcanic agglomerates between 1950 and 3000 m, the maximum depth reached (Fig. 5b). The volcanics are unconformably overlain by gas-bearing Valanginian sandstone. Details on the geological setting, exploration geophysics, structural evolution and reservoir characteristics are given in Perrot & Van der Poel (1987) and Herngreen et al. (1991). Magma may have reached the surface along Permo-Carboniferous and Kimmerian faults that were opened during the Late Jurassic. Based on geophysical data the agglomerates represent a neck, which forms part of a dome-like structure. This is illustrated by the isopach map of 'Upper Jurassic' units (Delfland Subgroup and Kimmeridge Clay), reflecting a division of the Vlieland Basin into two sub-basins and showing thinning around the dome (Fig. 5a). Perrot & Van der Poel (1987) interpreted local aeromagnetic anomalies in terms of a circular caldera-type body with a central volcanic plug (Fig. 5c). They hypothesised that, after a major eruption, the magma chamber must have collapsed, and that the volcanic islands were largely removed by Late Jurassic and Cretaceous erosion. The Wadden Volcaniclastic Member of the Delfland Subgroup, 78 m thick in *Slenk-1* and also found in *Zuidwal-2* and *Zuidwal-3*, consists of fine to coarse-grained volcaniclasts in a tuffaceous matrix. Because of their weathered appearance and proximity to the dome, they probably represent erosion products rather than primary volcanic deposits (Herngreen et al., 1991).

Radiometric dating of *Zuidwal-1* yielded variable results. Initial K-Ar ages obtained on four samples (1950-3000 m) ranged between 92 ± 2 and 119 ± 2 Ma, which, owing to extensive alteration, are probably minimum ages (Jeans et al., 1977; Harrison et al., 1979). The true age may be > 120 Ma, which was supported by a K-Ar date of 145 Ma obtained on a sample of unknown origin (personal communication G. Flacelière, in Jeans et al., 1977). Subsequent $^{40}Ar/^{39}Ar$ dating yielded a virtually identical age of 144 ± 1 Ma and a minor overprint between about 90 and 120 Ma, suggesting that the younger K-Ar ages may reflect argon loss (Dixon et al., 1981). Finally, Perrot & Van der Poel (1987) reported an age of 152 ± 3 Ma, based on the $^{40}Ar/^{36}Ar$–$^{40}K/^{36}Ar$ method.

Contrasting petrographic descriptions have been given. Dixon et al. (1981) infer that the eight altered rocks they examined originally were phonolite samples, biotite pyroxenites, a phonolitic basanite and a rock rich in pseudomorphed leucite. The agglomerate matrix attached to one of the phonolites contained leucite basanite and leucite tephrite clasts. Rock types referred to as trachytes, phonolites and leucitites from observations on 12 samples (Perrot & Van der Poel, 1987) are consistent with this description.

On the other hand, Harrison et al. (1979) described finely crystalline to partly glassy trachyte as the most common rock type, together with less abundant, heavily altered lava of probably basic composition and pieces of originally glassy vesicular pumice and minette. They noted that sphene is relatively abundant and do not record the presence of primary feldspathoids or their pseudomorphs. According to Dixon et al. (1981) it is conceivable that the Zuidwal volcanics represent one or more cycles of trachyte-phonolite eruptions and that they included subordinate amounts of more primitive lavas such as basanites and tephrites.

OTHER OCCURRENCES

Other Jurassic igneous rocks in the Dutch region have been found in the *E6-1* well on the rim of the Step Graben, with K-Ar ages of 161 ± 4 and 183 ± 4 Ma, and in *De Wijk-7* (2443-2486 m), east of the Texel-IJsselmeer High, with a K-Ar age of 155 ± 4 Ma. In both wells the rocks are unspecified intrusives in Carboniferous sediments (cf. Sissingh, 2004). Tuffaceous layers in the West Netherlands and Broad Fourteens basins (Fig. 2) have a stratigraphic age around the Jurassic-Cretaceous boundary.

The Dutch occurrences represent the southernmost expression of Jurassic magmatism in and around the North Sea (Fig. 6). Significant Middle Jurassic volcanism took place at the triple junction between the Moray Firth, the Viking Graben and the Central Graben in the northern North Sea, where an over 3000-m-thick sequence of basaltic lavas, the *'Forties Volcanics'*, formed a volcanic field (ca. 150-170 Ma) covering about 12 000 km² and extending eastward across the southernmost part of the Viking Graben (Woodhall & Knox, 1979). Age constraints, based

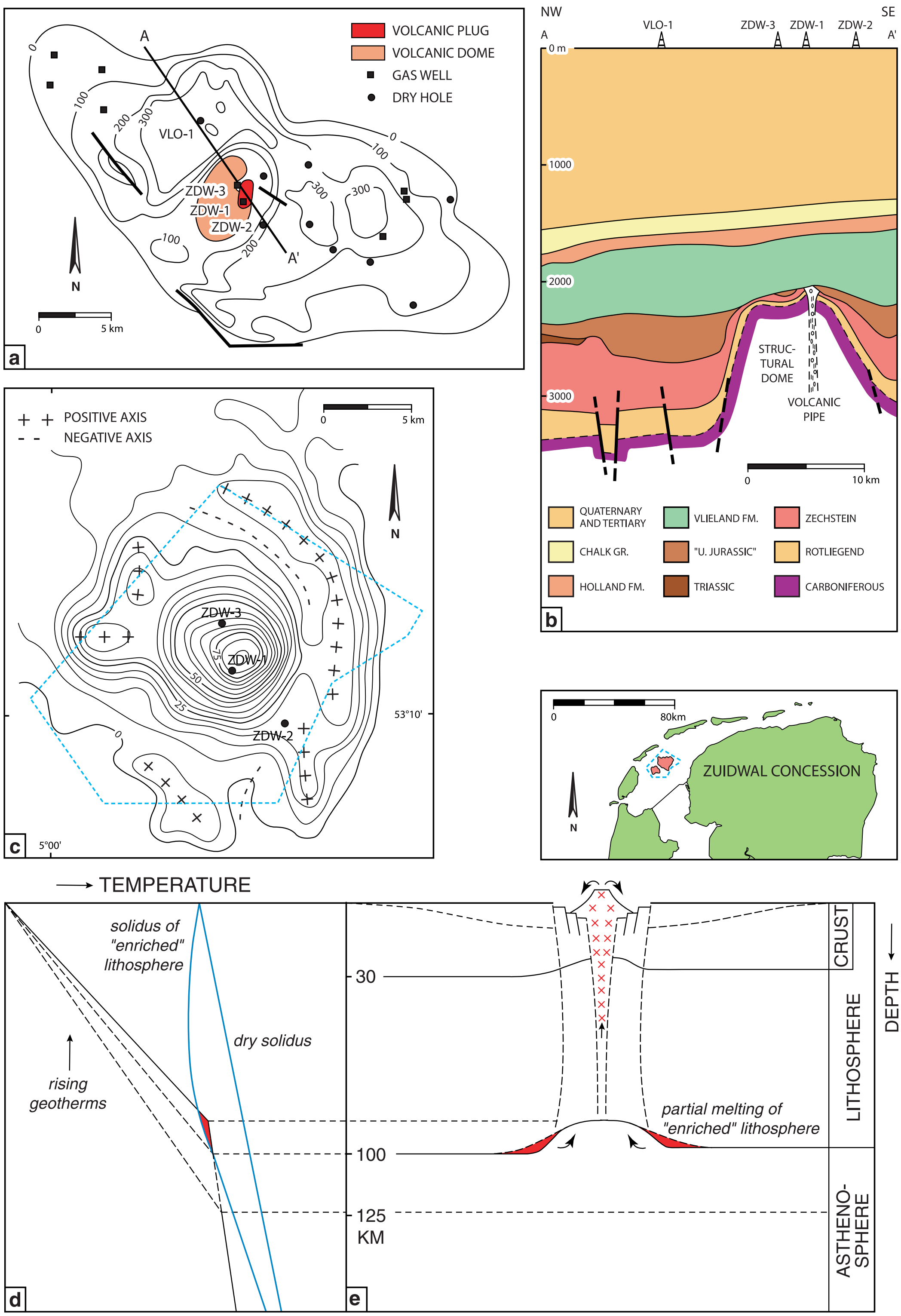

Fig. 5. Zuidwal Volcano (modified after Perrot & Van der Poel, 1987; Herngreen et al., 1991). (a) Isopach map (metres) of 'Upper Jurassic' in southern part of the Vlieland Basin. (c) Aeromagnetic map. See text for panels (d) and (e).

on predominantly K-Ar and some Ar-Ar data, have been discussed in Howitt et al. (1975), Ritchie et al. (1988) and Latin et al. (1990a), and stratigraphic relationships in Smith & Ritchie (1993). This episode may have succeeded earlier, Carboniferous and Middle Triassic volcanic activity

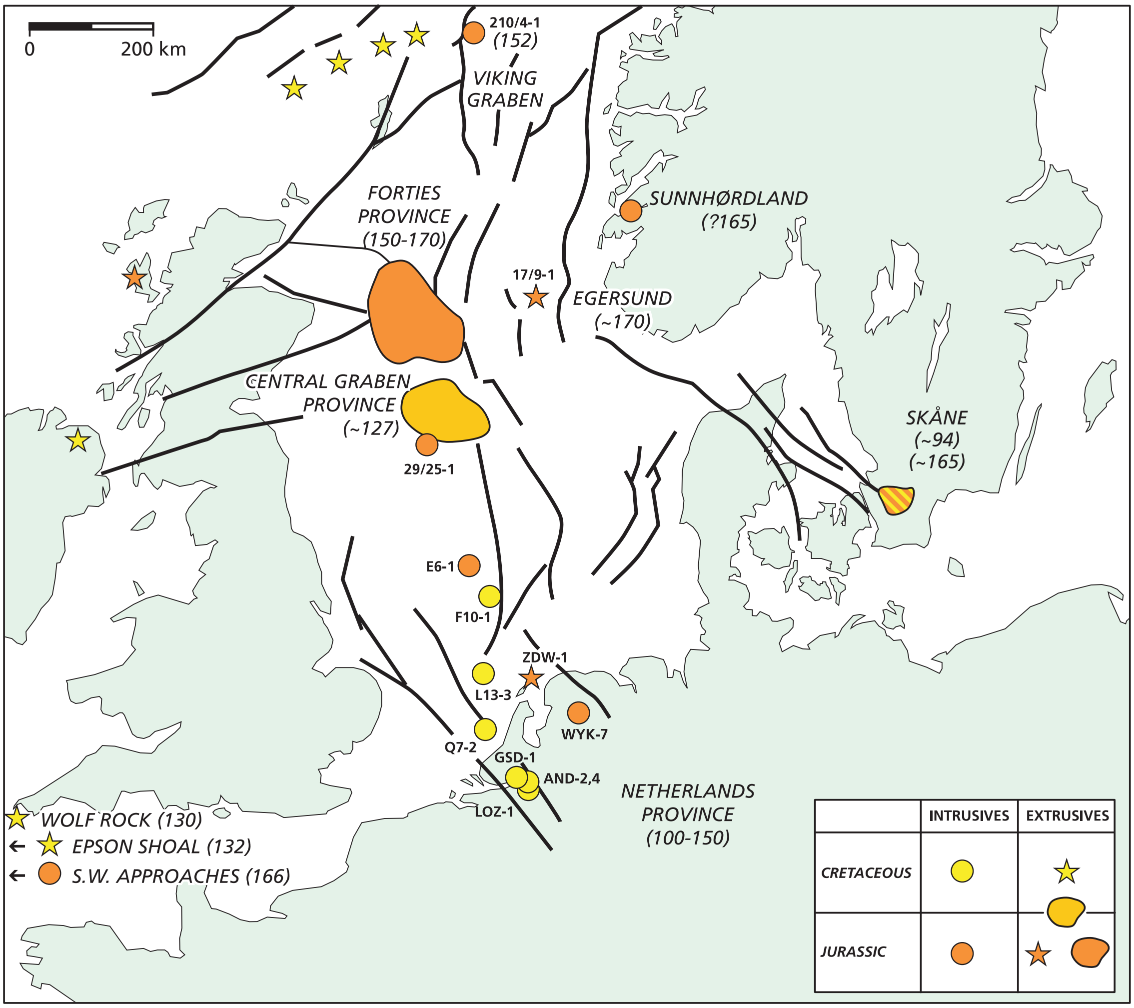

Fig. 6. Jurassic-Cretaceous igneous centres in NW Europe. Bracketed numbers refer to ages in Ma.

in the same area, and was followed by minor volcanism in the Early Cretaceous. Predominantly subaerial eruptions of modest explosivity are thought to have been fed from fissures rather than a central volcanic cone. Tuffs and agglomerates are subordinate in volume.

In general, the flows are mildly undersaturated alkali basalts or ankaramites, containing abundant phenocrysts of olivine and clinopyroxene. According to Gibb & Kanaris-Sotiriou (1976), the lavas have oceanic alkali-basalt affinities, whereas Dixon et al. (1981) pointed to a similarity with alkali-olivine basalt in continental rift settings. More recent major and trace-element data showed that all of the basalts belong to the alkaline series (Latin & Waters, 1992). Most are fairly primitive (MgO > 8%), whereas more evolved hawaiites and mugearites occur as well.

In the Norwegian *Egersund Basin*, in well 17/9-1, nephelinite lavas are interbedded with sediments, forming a several hundred metres thick sequence. K-Ar ages of 177-180 Ma (Furnes et al., 1982), a $^{40}Ar/^{39}Ar$ age of 170 ± 2 Ma (Latin & Waters, 1992) and a Bathonian to Bajocian age of the sediments make these lavas somewhat older than the main phase of Forties volcanism. They are porphyritic vesicular rocks with large clinopyroxenes and pseudomorphs after olivine, which are set in a fine-grained or glassy groundmass. High contents of incompatible trace elements confirm the alkaline nature of these rocks (Dixon et al., 1981; Latin & Waters, 1992). In the same well, a lower sequence of igneous rocks was cored that intruded Lower Jurassic and ?Triassic sediments (Latin et al., 1990a) and yielded a similar K-Ar age of 177-178 Ma (Furnes et al., 1982). These are strongly undersaturated mafic alkaline rocks with mineral assemblages and textures that resemble alnöites.

A group of Middle Jurassic alkali-olivine basalt flows and plugs in *Skåne* in south Sweden, comparable to the Egersund nephelinites, yielded whole-rock K-Ar ages of 163-173 Ma (references in Dixon et al., 1981). Other samples fall between 122 and 151 Ma, whereas a minor group of intrusions may have been emplaced during the mid-

Cretaceous (81-107 Ma). It should be noted, however, that Ar loss might have resulted in ages that are too young. A comprehensive petrological study showed that the Scanian volcanics are mainly basanites with more rare melanephelinites (Tappe, 2004). A Middle Jurassic olivine-biotite gabbro with a K-Ar age of 166 ± 4 Ma was encountered in a borehole in the Southwestern Approaches *off the west coast of Cornwall* (Harrison et al., 1979). Middle Jurassic smectitic clays in *southern and eastern England* have been interpreted as alteration products of volcanic air-fall deposits, part of which may originate from volcanic centres in the North Sea (Bradshaw, 1975; Jeans et al., 1977, 2000).

Cretaceous

Early Cretaceous undersaturated alkaline rocks occur in a number of wells in the West Netherlands Basin. The fine-grained igneous rocks in *Andel-2* and 4 and in *Loon op Zand-1* represent intrusives in Middle and Lower Jurassic sediments, respectively. A sample from Andel-4 has been $^{40}Ar/^{39}Ar$ dated at 133 ± 2 Ma and one from Loon-op-Zand-1 at 132 ± 3 Ma (Dixon et al., 1981). Because of a partial overprint at < 120 Ma the samples could be older (140-150 Ma?), perhaps similar in age to the Zuidwal volcanic centre. The Loon-op-Zand and Andel-4 rocks are highly altered, glassy olivine nephelinites, containing pseudomorphed olivine and clinopyroxene phenocrysts in a groundmass with the same mineral phases as well as kaersutitic hornblende, biotite and apatite. The Andel-2 samples are probably basanites with pseudomorphed olivine, plagioclase and possibly nepheline. In terms of immobile trace-element concentrations, this group of samples is comparable to the less altered nephelinites in the Egersund Basin (see below). Their ages correspond reasonably well with a K-Ar age of 125 ± 25 Ma obtained on an intrusive body of biotite-bearing(?) olivine-basaltic rock in the Portlandian-Valanginian(?) Alblasserdam sand-shale, sampled in *Giessendam-1* (Sissingh, 2004). Dixon et al. (1981) suggested that the *Berkel-1* hornblende basalt, intruded in Jurassic sediments (Van der Sijp, 1953), may be a fine-grained essexite or theralite, and possibly similar in age to the Andel and Loon-op-Zand rocks. This may also be the case for a hornblende-diabase intrusion in the lowermost Cretaceous of *Oldenzaal-2* in the eastern Netherlands, which has been described by Van Voorthuysen (1944).

Igneous rocks from Dutch offshore wells *F10-1 (=PL1)*, *K14-FA103*, *L13-3* and *Q7-2* are all strongly undersaturated basaltic intrusives (Dixon et al., 1981; Latin et al., 1990a). Except for K14-FA103, for which no age is available, they are clearly younger than the Early Cretaceous magmatic rocks found onshore in the West Netherlands Basin. According to phenocryst assemblages and bulk-rock data, the rocks from K14-FA103 (border area Broad Fourteens Basin) and F10-1 (near edge of Dutch Central Graben) share a potassium and volatile-rich character. The former constitute a thin fine-grained body of trachybasaltic composition in Lower Permian sandstone, whereas the latter are lamprophyric basanites, carrying abundant amphibole and biotite, that yielded a K-Ar age of 99 ± 5 Ma (Sissingh, 2004).

The other Dutch offshore occurrences are more nephelinitic, mafic intrusives, compositionally and petrographically comparable to the Andel and Loon-op-Zand rocks mentioned above. The rocks from well L13-3 that occur within Zechstein salts have a best apparent $^{40}Ar/^{39}Ar$ age of 101 ± 1 Ma showing a partial overprint at 51.9 ± 0.3 Ma. A dated sample from Q7-2 represents an igneous body intruded into Triassic rocks, and has an irregular $^{40}Ar/^{39}Ar$ spectrum yielding a most probable age of between 95 ± 2 and 106 ± 2 Ma (Dixon et al., 1981).

Outside the Netherlands, Early Cretaceous igneous rocks have been found in wells in the UK part of the *Central Graben* area (Fig. 6). Two sequences of mafic biotite-phonolite intrusive rock, interpreted as a single dyke or two sills, occur in Zechstein deposits in well 29/25-1 on the edge of the Mid North Sea High. $^{40}Ar/^{39}Ar$ dating yielded 138 ± 4 Ma for the upper sequence (Dixon et al., 1981). Smith & Ritchie (1993), however, argued that this age is erroneous and that these rocks are associated with an early Mid Jurassic volcanic centre. A lava flow cored in a well in the Auk oil field (*30/16-A11*) has a K-Ar age of 127.4 ± 2.6 Ma. An undersaturated and potassium-rich character and mineral assemblages including primary biotite and amphibole are common features of these Central Graben rocks (Latin et al., 1990a). Phonolitic rocks of comparable K-Ar age have been identified on the Isles of Scilly to the west of Cornwall (references in Harrison et al., 1979). The sodium-rich *Wolf Rock* (130 ± 5 Ma) represents a feeder or stump of a volcanic neck, whereas the shape of *Epson Shoal* (132 Ma) is less clear. Jeans et al. (2000) referred to the igneous centres in the West Netherlands Basin and other centres in the North Sea area as a potential source of argillized volcanic ash deposits in the Lower Cretaceous of England. The Wolf Rock and Epson Shoal centres fall in the same age range. Altered volcanic ashes also occur in several younger deposits in England, up into the Upper Cretaceous. Evidence for Aptian igneous activity in the Lower Saxony Basin comes from the occurrence of tuffs near Hannover and from the presence of deep-seated laccolithic intrusions (e.g. Bramsche Massif) inferred from geophysical anomalies and coalification of overlying sediments (Ziegler, 1990; Brink et al., 1992, and references therein; but see Senglaub et al., 2006, for an alternative interpretation of the Bramsche anomaly).

Late Cretaceous igneous rocks are restricted to basalts and gabbroic intrusives in *Ireland* and in the *Rockall-Faeroe Trough area* in the Atlantic Ocean north and west of the

British Isles (Ziegler, 1990). They have K-Ar ages between 70 and 81 Ma. Volcanogenic sediments have been identified in the Turonian and Maastrichtian of north Germany (references in Harrison et al., 1979). Volcanic ash may also have been deposited in the Netherlands during the Late Cretaceous in view of the presence of the tuffaceous layers in England and Germany, but the small grain size and easy alteration make the identification of minor volumes of air fall uncertain.

Early Tertiary

Ash layers identified from petrophysical information occur in the Eocene Dongen Formation. The southern limit of these deposits is shown in Figure 2. Equivalent deposits are widespread in the offshore and onshore stratigraphic records of the UK, Germany and Denmark, and are the expression of an episode of explosive subaerial volcanism. A regionally distributed earliest Eocene ash deposit represents a conspicuous marker in the North Sea Basin (Jacqué & Thouvenin, 1975). The eruptive centres were presumably associated with the large North Atlantic Thulean volcanic province, where major intrusive and extrusive magmatism with a bimodal character took place in the same period (Ziegler, 1990; Ritchie et al., 1999). Explosive volcanism in the Skagerrak area has been considered as an alternative source of Eocene ash deposits (cf. Nielsen & Heilmann-Clausen, 1988, and references therein) but the magnetic anomaly on which this hypothesis was based probably signals a Sveconorwegian intrusion in the basement rather than a younger magmatic body (Olesen et al., 2004).

The Lower Oligocene (Rupelian) Boom or Rupel Clay, present in the subsurface in much of the Netherlands, carries evidence of volcanic activity in exposures in northern Belgium. A detailed petrographic study has shown that the 50 to 75-m-thick marine shelf sediment contains a considerable fraction of volcanic material (Zimmerle, 1993). A significant portion of the clayey matrix is thought to consist of the alteration products of primary clay-size volcanic ash particles of trachytic and basaltic composition. An admixture of coarser-grained particles of mineral phases and argillized lithoclasts also point to a volcanic provenance. Plausible sources are the Siebengebirge and the Hocheifel, volcanic centres situated south of the Lower Rhine Embayment that were active from the Eocene to the Miocene according to K-Ar ages summarised in Zimmerle (1993). Direct ash fall, erosion of volcanic soils and longshore volcanic drift were possible modes of transport. The products of the Siebengebirge represent an alkali basalt-trachyte association. The trachytes include explosive varieties and pyroclastic flow deposits that are possibly associated with caldera-forming events (Vieten et al., 1988). The Hocheifel has also produced lavas and pyroclastics with an alkali-basaltic affinity (Huckenholz & Büchel, 1988).

Quaternary

Distal tephra layers have been identified in Lateglacial sediments at Kostverloren Veen (province of Drenthe), one of the most north-easterly pingo remnants in the Netherlands (Davies et al., 2005). Geochemical fingerprints identified glass shards of one layer as the rhyolitic version of the Vedde Ash (mid-Younger Dryas), an important regional stratigraphic marker in the North Atlantic, the Norwegian Sea, and the adjacent land area. This occurrence of the rhyolitic Vedde Ash belongs to one of two main plumes, thought to originate from southern Iceland, that extends in an easterly/south-easterly direction over northern Britain, southern Scandinavia and western Russia.

Indirect evidence for magmatic intrusions

There is indirect evidence for the presence of two sizeable intrusive bodies along the border with Germany (Fig. 2). The East Groningen Massif is defined by a coalification anomaly measured in Upper Carboniferous sediments from wells in the Ems estuary in north-eastern Groningen and in the adjacent part of Germany (Kettel, 1983). The shape of the anomaly coincides with a positive magnetic anomaly and with structural contours of the top of the Rotliegend for Jurassic times. This, in combination with the ages of the Zuidwal, Andel and Loon-op-Zand igneous occurrences, led Kettel (1983) to assume that the East Groningen Massif is related to an intrusive body that was emplaced around the Jurassic-Cretaceous boundary. It cannot be excluded, however, that the inferred intrusion is older and possibly Early Permian in age, as suggested by Van Wijhe et al. (1980).

The location of the Erkelenz intrusion in the province of Limburg and the adjacent part of Germany has been inferred from geophysical surveys. A pronounced magnetic anomaly, in combination with a modest gravimetric anomaly, may point to an acid rather than a basic body (Bredewout, 1989). There are no data for the timing of this intrusion, but the heat pulse is reflected in the degree of coalification of overlying Upper Carboniferous strata (Teichmüller & Teichmüller, 1971). A Permian age would be consistent with the widespread presence of other manifestations of acidic magmatism in north-west Europe, in contrast with the predominantly basaltic nature of Jurassic igneous rocks.

The pre-Permian residual gravity field, obtained after subtracting the contribution of the younger sedimentary succession to the total gravity field, shows an anomaly pattern for the Dutch onshore and offshore region that reflects sources in the crust or at the crust-mantle boundary. If the former option applies, positive residual anomalies can be explained by high-density magmatic intrusions at pre-Permian levels (Dirkzwager et al., 2000).

Interestingly, large parts of the positive areas coincide with the Dutch Central Graben and the West Netherlands and Broad Fourteens basins, areas where intrusive and extrusive rocks have been found in a fair number of wells.

Geochemical signatures

Petrographic observations show that secondary alteration is a ubiquitous feature in virtually all of the Paleozoic and Mezosoic igneous rocks that were encountered in the wells mentioned above. Hence, the analytical data of bulk-rock samples generally reflect modifications of original geochemical signatures to a certain extent. For this reason, magmagenetic interpretations often rely on 'immobile' minor and trace elements that are considered to be relatively insensitive to alteration processes. Widely used examples are Ti, P, Zr, Nb, Y and the rare-earth elements (REE). Partly because of the alteration problem, geochemical data are not available for most of the Dutch igneous rocks, particularly those older than Cretaceous. Table 2 lists published data for rocks from wells in the Netherlands and adjacent north-west European areas.

According to petrographic descriptions only, the bimodal character of the Permian rocks in Dutch wells, expressed by their basaltic and rhyolitic composition, is also seen in the Danish offshore, whereas the German Rotliegend occurrences show a wide compositional variation. Examples are given in Table 2. As noted by Dixon et al. (1981) the basalts from the Danish North Sea region are mildly alkaline or transitional in having an alkali-basalt mineralogy but an olivine-tholeiite normative chemistry. In a total-alkali versus SiO_2 diagram, they plot in the upper part of the basalt field (Fig. 7a). The ca. 299 Ma old samples from well 39/2-4 on the western flank of the UK Central Graben are tholeiitic basalts with low abundances of incompatible trace elements, including the light rare-earth elements (REE), and distinct trace-element ra-

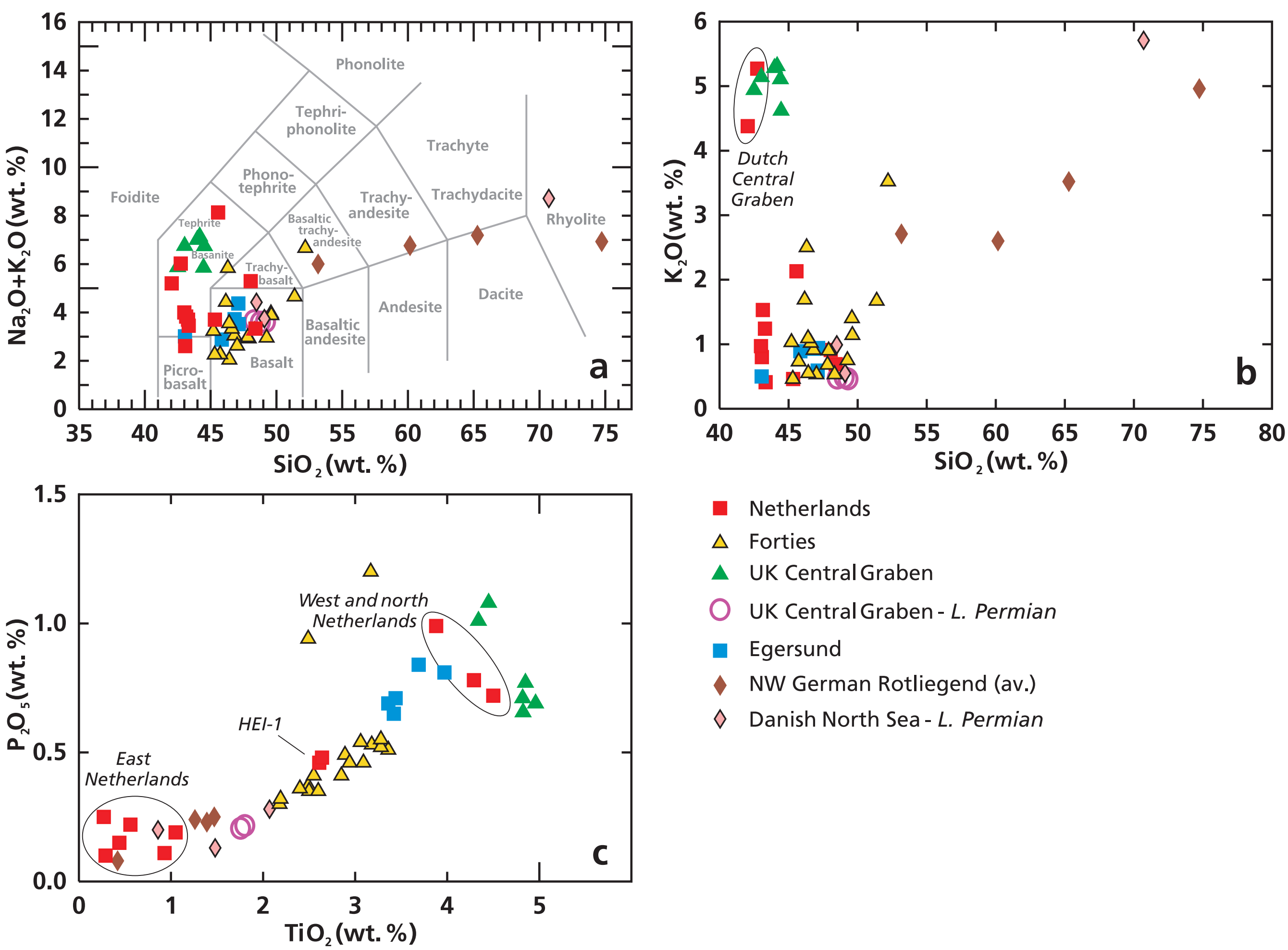

Fig. 7. Bulk-rock compositions of igneous rocks in the Netherlands and adjacent areas (data in Table 2). (a) SiO_2-alkalies classification diagram. (b) SiO_2-K_2O diagram. (c) TiO_2-P_2O_5 diagram.

Table 2. Geochemical data of igneous rocks in the Netherlands and adjacent areas (continued next page).

Area Age	Netherlands Carboniferous-Cretaceous											Egersund Basin, Norway Jurassic				
Well/location	F10-1 (=PL-1)		Andel-2	Wanneperveen		De Wijk-7	Dwingelo	Corle	Hupsel	Heinenoord-1		17/9-1	17/9-1	17/9-1	17/9-1	17/9-1
Code	PL1/10	PL1/12	G-Andel2	WAV (2070 m)	WAV (2015 m)	WYK-7 (2690 m)	DWL (3797 m)	COR (970 m)	GEL-3 (1320 m)	HEI-1 (2236.5 m)	HEI-1 (2238.5 m)	7N	ES1/24	ES1/34	D-17/9-1	8N
Rock type	Ultrapotassic lamprophyre	Ultrapotassic lamprophyre	Olivine nephelinite/ basanite	Olivine gabbro	Olivine gabbro	Olivine gabbro	Olivine gabbro	Melaphyre	Leucophyre	Alkali basalt	Alkali basalt	Sodalite nephelinite	Sodalite nephelinite	Sodalite nephelinite	Nephelinite	Sodalite olivine nephelinite
Reference	1	1, 2	2	3	3	3	3	3	3	4	4	1	1	1	2	5
Major elements (wt.%)																
SiO_2	40.44	40.13	43.44	40.87	40.15	45.99	41.32	46.38	37.61	41.18	41.74	43.11	45.37	44.38	42.68	47.00
TiO_2	4.06	4.30	3.91	1.00	0.49	0.42	0.26	0.90	0.24	2.48	2.55	3.16	3.31	3.76	3.66	3.41
Al_2O_3	14.59	14.45	18.68	10.45	16.23	16.14	11.07	15.44	18.95	12.34	12.79	11.43	11.71	13.00	12.44	12.29
Fe_2O_3[1]			14.06	7.21	3.79	4.81	3.70	4.69	3.82	10.84	11.00				14.13	10.28
FeO[1]	11.17	11.17		9.11	7.84	6.52	8.79	8.03	9.69			8.95	8.69	9.39		
MnO	0.24	0.24	0.12	*trace*						0.17	0.17	0.53	0.26	0.20	0.21	0.16
MgO	10.51	10.74	7.74	15.31	8.25	8.11	19.88	7.38	4.22	12.57	12.22	6.68	8.11	7.77	10.07	7.55
CaO	7.20	8.78	9.27	7.12	3.99	9.70	6.78	8.49	5.29	11.59	12.12	16.81	14.74	12.00	12.14	14.07
Na_2O	0.71	0.78	1.83	2.88	5.29	2.51	2.90	4.23	2.69	2.34	2.23	1.87	2.48	2.64	2.49	3.77
K_2O	4.99	4.18	0.81	0.92	1.88	0.66	0.39	0.88	0.38	1.18	1.48	0.84	0.9	0.88	0.5	0.59
P_2O_5	0.74	0.69	1.00	0.18	0.19	0.14	0.24	0.11	0.08	0.44	0.46	0.65	0.68	0.77	0.83	0.65
Total	94.65	95.46	100.86	95.05	88.10	95.00	95.33	96.53	82.97	95.12	96.76	94.03	96.25	94.79	99.15	99.77
H_2O^+				2.77	2.94	2.31	3.00	2.02	13.03							
CO_2				1.80	2.77	0.94	1.56	0.70	3.90							
LOI	4.0	3.20	9.80							3.07	2.43	4.0	2.8	4.0	6.30	5.12
Trace elements (ppm)																
V	459	501	439									525	402	485	451	
Ba	814	760	905									433	352	4751	293	
Sc	26	36	38									29	36	37	42	
Cr	19	32	312									266	220	189	245	
Ni	34	46	151									72	9	6	63	
Cu	90	69	58									37	51	114	110	
Zn	80	94	79									72	133	123	96	
Rb	61	63	24							30	40	44	46	109	30	26
Sr	1174	893	823							1360	970	648	477	616	519	379
Y	27	27	24							20	20	21	19	24	23	16
Zr	321	318	311							150	160	331	316	360	354	337
Nb	103	94	109									129	120	140	154	132
La	82	74	92									72	77	72	95	
Ce	157	150	157									144	156	152	208	
Pr	16	17										16	17	17		
Nd	63	67	57									60	65	66	70	
Sm	10.0	10.6										9.0	9.5	10		
Eu	3.0	3.2										2.5	2.6	3.0		
Gd	8.4	8.9										6.9	7.3	8.2		
Tb																
Dy	6.2	6.3										4.6	4.8	5.5		
Ho	1.1	1.1										0.8	0.8	0.9		
Er	2.7	2.9										2.0	2.1	2.4		
Tm																
Yb	2.1	2.1										1.5	1.5	1.8		
Lu	0.3	0.3										0.2	0.2	0.3		
Pb		2	4							<10	<10				5	
Th		6	8							<10	15				10	
Co																

Table 2. Continued.

Area	Forties Volcanics Field, North Sea																	
Age	Jurassic																	
Well/location	15/21-8a	15/21-8a	AH1	AH1	21/9-1	21/9-1	21/9-1	21/10-1	21/9-1	21/3-2	21/3-2	15/27-2	15/27-2	21/9-1	21/9-1	21/10-1	21/9-1	21/9-1
Code	AH1/36	AH1/42	A-AH1	B-AH1	BP1/1	BP1/33	BP1/64	BP2/1	C-21/9-1(BP)	TO1/21	TO1/27	FO1/3	FO1/4	9.C	9.B	11211	3395.25	3396.5
Rock type	Alkali basalt	Alkali basalt	Ankaramitic alkali basalt	Altered ankaramitic alkali basalt	Alkali basalt	Alkali basalt	Alkali basalt	Alkali basalt	Alkali basalt	Alkali basalt	Alkali basalt	Bas. trachy-andesite	Alkali basalt	Alkali olivine basalt	Alkali olivine basalt	Alkali olivine basalt	Alkali olivine basalt	Alkali olivine basalt
Reference	1	1	2	2	1	1	1	1	2	1	1	1	1	7	7	7	7	7
Major elements (wt.%)																		
SiO_2	47.74	45.11	47.81	50.72	45.44	45.98	46.65	44.95	45.00	47.7	48.38	50.39	44.65	44.61	43.96	44.27	44.2	43.59
TiO_2	2.11	2.40	2.19	3.32	2.99	2.74	2.95	2.42	3.26	2.31	2.54	2.40	3.04	2.78	3.12	2.47	2.87	2.99
Al_2O_3	11.82	13.28	11.90	17.63	13.43	12.96	14.02	9.89	15.07	12.77	12.68	18.48	17.65	12.84	13.95	9.72	12.51	13.39
Fe_2O_3 [1]			12.51	7.97					11.97					6.91	7.13	4.44	6.09	8.85
FeO [1]	8.88	9.57			8.4	9.42	7.99	8.1		8.86	9.01	6.96	8.85	4.45	4.53	5.84	5.43	3.46
MnO	0.14	0.20	0.11	0.11	0.41	0.25	0.29	0.21	0.40	0.23	0.2	*n.d.*	*n.d.*	0.22	0.22	0.20	0.40	0.14
MgO	9.65	10.71	9.63	2.57	10.53	9.85	9.32	15.32	8.99	10.07	10.04	4.40	4.90	10.6	9.18	14.27	11.26	8.66
CaO	13.42	11.81	12.43	11.30	12.46	11.71	11.65	13.61	11.10	10.09	10.54	6.56	12.37	9.84	8.40	13.01	12.13	7.07
Na_2O	2.13	2.05	2.02	2.94	2.02	2.20	2.63	1.44	2.19	2.45	2.67	3.03	2.20	2.36	2.61	1.48	1.75	3.13
K_2O	0.73	0.88	0.90	1.65	0.51	0.65	0.51	0.53	1.03	1.35	1.11	3.40	0.98	1.05	1.61	0.71	0.45	2.35
P_2O_5	0.29	0.36	0.32	0.5	0.44	0.39	0.52	0.34	0.52	0.35	0.34	0.91	1.15	0.47	0.52	0.40	0.45	0.50
Total	96.91	96.37	99.82	98.71	96.63	96.15	96.53	96.81	99.53	96.18	97.51	96.53	95.79	96.13	95.23	96.81	97.54	94.13
H_2O^+														3.55	4.47	2.87	3.06	5.52
CO_2														0.22	0.28	0.22		0.28
LOI	1.1	1.5	2.60	10.20	2.9	2.5	2.7	3.3	2.90	2.7	3.0	2.6	2.7					
Trace elements (ppm)																		
V	363	383	396	571	379	313	338	267	360	290	312	259	321					
Ba	557	511	502	1022	596	573	466	497	593	689	521	1438	3534					
Sc	46	44	50	57	37	35	31	38	32	28	37	5	7					
Cr	766	277	815	336	291	359	283	1171	132	287	464	2	2					
Ni	241	127	236	189	110	152	107	83	65	163	167	69	9					
Cu	34	63	57	56	39	39	42	61	72	57	93	9	21					
Zn	77	85	107	36	113	84	80	84	108	93	85	165	130					
Rb	14	6	32	64	3	4	3	15	15	36	36	94	15					
Sr	471	482	481	705	746	698	822	316	905	774	919	1344	1653					
Y	22	24	21	22	27	27	29	21	30	26	6	34	35					
Zr	163	197	166	255	240	221	282	193	273	208	190	424	386					
Nb	36	46	37	61	64	56	65	52	78	51	48	147	126					
La	26	35	29	41	44	39	46	38	56	39	35	80	93					
Ce	56	73	64	81	94	84	101	79	130	80	72	166	187					
Pr	6.8	8.7			11	9.9	12	9.2		9.0	8.4	18	21					
Nd	30	36	27	34	45	41	49	38	56	37	35	72	82					
Sm	5.7	6.7			7.8	7.3	8.5	6.3		6.5	6.3	11.8	13.1					
Eu	1.75	2.04			2.49	2.31	2.61	1.93		1.96	1.96	3.62	3.89					
Gd	5.3	6.3			6.9	6.5	7.2	5.4		5.8	5.8	9.7	10.8					
Tb																		
Dy	4.5	5.2			5.3	5.2	5.6	3.9		4.7	4.6	7.2	8.0					
Ho	0.77	0.9			0.9	0.89	0.97	0.65		0.82	0.8	1.22	1.35					
Er	2.11	2.45			2.39	2.42	2.61	1.69		2.22	2.13	3.26	3.57					
Tm																		
Yb	1.67	1.97			1.82	1.88	2.02	1.24		1.81	1.67	2.53	2.68					
Lu	0.24	0.29			0.26	0.27	0.29	0.17		0.27	0.26	0.37	0.38					
Pb									1									
Th			1	5					2									
Co																		

Table 2. Continued.

Area	UK Central Graben, North Sea									Horn Graben, DK		Ringkøbing-Fyn High, DK	NW Germany			
Age	Cretaceous						L. Permian			L. Permian		L. Permian	Rotliegend			
Well/location	30/16-A13Y	30/16-A13Y	30/16-A13Y	30/16-A13Y	29/25-1	29/25-1	39/2-4	39/2-4	39/2-4	R-1	R-1	D-1				
Code	SH1/5	SH1/10	SH2/2	SH2/3	8U	E-29/25-1	9115-0	9117-0	9125-5	18D	19D	10D	RTL-bas	RTL-and	RTL-dac	RTL-rhy
Rock type	Ultrapotassic extrusive	Ultrapotassic extrusive	Ultrapotassic extrusive	Ultrapotassic extrusive	Mafic biotite phonolite (monchiquite)	Monchiquite	Tholeiitic basalt	Tholeiitic basalt	Tholeiitic basalt	Basalt	Rhyolite or silicified trachyte	Basalt	Average of 9 basalts	Average of 62 andesites	Average of 19 dacites	Average of 15 rhyolites
Reference	1	1	1	1	5	2	6	6	6	5	5	5	8	8	8	8
Major elements (wt.%)																
SiO_2	40.79	42.81	42.77	41.78	44.78	43.62	49.31	48.46	48.95	48.68	71.25	49.08	50.82	58.12	64.00	74.43
TiO_2	4.27	4.18	4.72	4.68	4.85	4.9	1.75	1.80	1.75	2.08	0.87	1.48	1.40	1.34	1.24	0.41
Al_2O_3	13.88	13.4	16.86	16.92	17.14	17.14	17.78	17.70	17.94	16.49	12.72	16.84	16.62	15.80	15.02	11.93
Fe_2O_3[1]					12.61	12.51	10.16	11.36	10.68	12.79	5.57	11.12	8.26	5.60	5.65	3.78
FeO[1]	12.85	11.45	8.72	11.69									4.25	3.49	1.94	0.4
MnO	0.28	0.48	0.14	0.17	0.11		0.09	0.10	0.10	0.26	0.12	0.13	0.18	0.17	0.10	0.04
MgO	9.83	8.47	8.51	8.97	8.2	7.49	7.42	6.84	7.70	6.47	0.29	8.78	5.25	4.39	2.18	1.18
CaO	7.4	8.9	7.98	5.65	5.52	5.36	9.56	9.51	9.12	8.92	0.98	8.67	7.25	4.60	2.65	0.88
Na_2O	0.89	1.18	1.67	1.56	1.71	1.81	3.09	3.15	3.10	3.44	3.02	3.19	3.14	4.02	3.60	1.96
K_2O	4.74	4.45	5.13	4.99	5.14	5.24	0.50	0.52	0.50	0.99	5.75	0.55	2.59	2.51	3.45	4.94
P_2O_5	1.04	0.97	0.75	0.69	0.66	0.68	0.21	0.22	0.21	0.28	0.2	0.13	0.24	0.22	0.24	0.08
Total	95.97	96.29	97.25	97.10	100.72	98.75	99.87	99.66	100.05	100.40	100.77	99.97	100.00	100.26	100.07	100.03
H_2O^+																
CO_2																
LOI	3.8	2.5	11.9	9.0	12.07	9.4	1.80	1.42	2.20	3.32	1.08	6.03				
Trace elements (ppm)																
V	414	404	503	509		523	253	247	253				171	127	80	20
Ba	886	1046	966	980		1040	227	226	226				397	563	521	706
Sc	24	26	46	49		49	34	34	35				37	31	27	9
Cr	4	*n.d.*	63	66		69	74	69	72				121	93	37	13
Ni	15	15	56	55		98	149	152	161				24	27	11	13
Cu	16	32	48	95		32	48	42	46				22	35	31	33
Zn	152	157	97	167		166							115	189	130	35
Rb	76	74	87	98	80	97	2.8	2.5	2.7	21	162	53	113	107	161	215
Sr	1083	1164	943	1022	830	1043	451	463	460	424	121	468	160	229	132	88
Y	32	32	25	23	14	22	22	23	22	21	66	18	33	37	51	58
Zr	402	387	339	334	341	360	93	94	94	176	738	85	182	196	294	368
Nb	111	109	104	98	108	106	7.3	7.6	7.4	10	45	11	15	18	27	28
La	89	85	64	69		84	8.2	8.2	7.5				26	36	58	63
Ce	186	177	132	143		171	16.2	18.2	17.2				58	71	100	109
Pr	21	20	15	16			2.5	2.8	2.8							
Nd	82	79	57	63		64	11	11	11							
Sm	13	13	8.8	9.7			2.9	2.8	2.8							
Eu	3.9	3.7	2.6	2.8			1.3	1.6	1.6							
Gd	11	10	7.3	7.7			3.6	3.6	3.7							
Tb							0.6	0.7	0.7							
Dy	8.0	7.7	5.2	5.4			3.4	4.0	3.6							
Ho	1.4	1.3	0.9	0.9			0.8	0.8	0.8							
Er	3.6	3.5	2.3	2.4			1.9	2.2	1.8							
Tm							0.3	0.3	0.3							
Yb	2.7	2.5	1.7	1.8			1.6	1.9	1.6							
Lu	0.4	0.4	0.2	0.3			0.2	0.3	0.3							
Pb						14							24	56	31	122
Th						8	0.4	0.2	0.2				10	11	16	21
Co													43	38	23	85

[1]All Fe is expressed as FeO or Fe_2O_3, except when data for both oxides are given. LOI: loss on ignition in wt.%. Blank fields: no data reported. References: 1: Latin & Waters (1992), 2: Latin et al. (1990a), 3: Eigenfeld & Eigenfeld (1986), 4: Helmers (1991), 5: Dixon et al. (1981), 6: Heeremans et al. (2004), 7: Gibb & Kanaris-Sotiriou (1976), 8: Eckhardt (1979). DK: Denmark, RTL: Rotliegend.

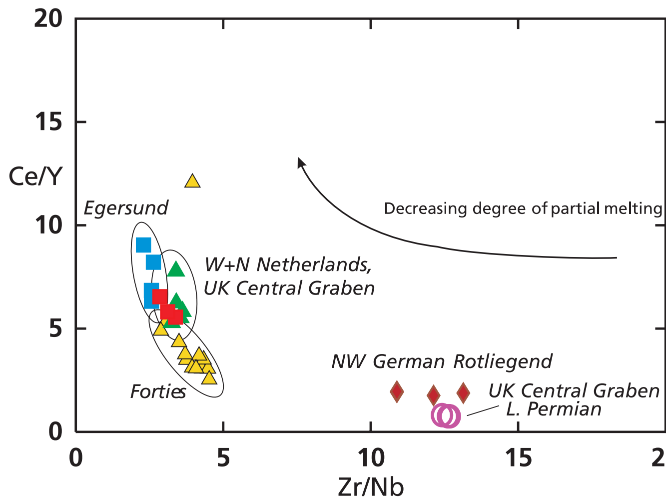

Fig. 8. Diagram of Zr/Nb versus Ce/Y ratios, illustrating differences in inferred degrees of partial melting between magmatic centres in different parts of NW Europe (simplified after Latin et al., 1990a). Symbols as in Fig. 7; data in Table 2.

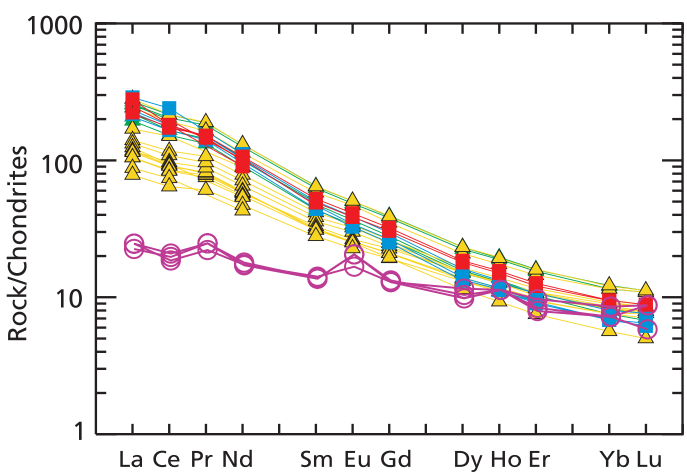

Fig. 9. REE patterns of Jurassic-Cretaceous igneous rocks in NW Europe. Symbols as in Fig. 7; data in Table 2.

tios in comparison to the Jurassic and Cretaceous magmatic rocks in the North Sea (Figs 8, 9). They deviate from normal mid-ocean-ridge basalts (N-MORB) in being more enriched in light and more depleted in heavy REE (Heeremans et al., 2004). Immobile-trace-element signatures confirm a 'within-plate' tectonic affinity for both the Danish and the UK rocks (Fig. 10). On the other hand, basalts in the North-east German Basin show a wide diversity in tectonic discrimination diagrams (Benek et al., 1996).

Analytical data of a large number of 'spilitised' volcanics in the north-west German Ems-Weser area show a dominance of intermediate compositions and a wide range between basaltic and rhyolitic rock types (Eckhardt, 1968, 1979). The least evolved rocks have an alkaline affinity, straddling the boundary between the basalt and trachybasalt fields. Averages for all of these German Rotliegend volcanics confirm the compositional range and the alkali enrichment that characterises this magmatic episode (Figs 7a, b).

The Jurassic-Cretaceous rocks are the best documented in terms of geochemistry, although still relatively few data have been reported for the Dutch occurrences (Dixon et al., 1981; Eigenfeld & Eigenfeld, 1986; Latin et al., 1990a; Helmers, 1991), partly due to the strong alteration that is generally observed. Although all are undersaturated, interesting variations can be recognised. The most noticeable is the difference between the lamprophyric rocks of the F10-1 well and the nephelinite of the Andel well. Both groups are basaltic, but the former are richer in potassium, and total alkalies, and similar to the UK Central Graben rocks, whereas the latter contains less potassium and resembles the Egersund rocks (Table 2, Figs 7a, b). In terms of immobile minor and trace elements, most of the west and north Netherlands rocks overlap the field of the UK Central Graben rocks and are distinct from the Forties and Egersund fields (Latin et al., 1990a, b; Figs 7c, 8). This difference also appears in the REE patterns, which show almost straight lines and no Eu anomalies (Fig. 9). In all cases the light rare-earth elements are enriched over the heavy, but to a different extent, as Ce/Yb ratios are highest in the Egersund rocks, intermediate in the Netherlands + UK Central Graben group and lowest in the Forties basalts. With Ce/Y and Nb/Zr ratios decreasing in the same order (Fig. 8), these trace-element signatures may reflect an increasing degree of partial melting of a common source (Latin et al., 1990a). Such systematics cannot be evaluated for the Heinenoord-1 alkali basalt in the West Netherlands Basin and the gabbroic rocks in the east Netherlands, owing to the absence of sufficient trace-element data. Interestingly, in the TiO_2-P_2O_5 diagram (Fig. 7c) the Heinenoord-1 rock is distinct from the other west and north Netherlands samples, and plots in the Forties field, whereas the east Netherlands intrusives show similarities with the Danish and German Permian volcanics. The mafic rocks from all of these regions plot in the 'within-plate' field in a Ti-Zr-Y tectonic discrimination diagram (Fig. 10), with the exception of the German Rotliegend averages of Eckhardt (1979), which fall in the calcalkali field. As Benek et al. (1996) have shown for the Permo-Carboniferous volcanics in the North-east German Basin, the geochemistry of the Rotliegend rocks may show spatial variations and deviates from a typical within-plate signature according to the extent to which magma sources had been affected by Variscan tectonics.

For the highly altered Zuidwal volcanics a complete set of bulk-rock data has not been published. However, the composition of relict pyroxenes is typical of highly undersaturated feldspathoidal magmas (Dixon et al., 1981), and immobile-element signatures of the least evolved samples indicate that they also belong to the Netherlands + UK Central Graben group, intermediate between the Forties alkali basalts which have no modal feldspathoids and the Egersund rocks which have no modal feldspar (Latin et al., 1990a). The only Sr-Nd-Pb isotopic data available are for bulk rock and clinopyroxene separates from the wells F10-1 (Dutch Central Graben), 15/21-8a (Forties) and 17/9-1 (Egersund) (Latin & Waters, 1992). The initial isotope ratios are similar, and tend to be somewhat more 'enriched' as compared to values for a mid-ocean-ridge basalt (MORB) mantle source.

Magmagenesis and rifting

Jurassic-Cretaceous

The relation between Jurassic-Cretaceous magmatism and rifting in the Netherlands is best illustrated in conjunction with the magmatic occurrences in other parts of the North Sea Basin. Differences in timing, location, volume and composition allow constructing a coherent genetic framework for the entire region (Latin et al., 1990a, b; Latin & Waters, 1992). The Forties basaltic province, which can be seen as the focal point of the magmatic activity, is situated within the main rift system. The much less voluminous occurrences of the Netherlands, North Sea Central Graben and Egersund Basin are restricted to the flanks of the main rift system or to minor sub-basins. All of these magmas were produced during the syn-rift phase of basin development, which probably lasted from ca. 250 until ca. 100 Ma (Ziegler, 1990; White & Latin, 1993), but the igneous activity tends to become progressively younger in southward direction from Mid Jurassic to Early Cretaceous, perhaps reflecting a propagation of the rift system (Dixon et al., 1981; Latin et al., 1990a) and associated activation of the West Netherlands and Vlieland basins (cf. Ziegler, 1990).

None of the magmas is derived from an asthenospheric mantle source similar to that producing mid-ocean-ridge basalts. The Forties basalts represent the largest-degree melts in the Mesozoic (< 2%), and did originate from asthenospheric mantle, but their trace-element and isotope signatures indicate that an enriched component is involved as well, either as heterogeneities in the asthenosphere or in the form of lithospheric melt that mixed with the asthenospheric melt. The other magmas, including those of the Netherlands, must have formed by lower-degree melting of a volatile- and incompatible-trace-element-enriched region of the mantle, which had remained separated from the asthenosphere for hundreds of millions of years. Therefore, the continental lithosphere is the most plausible source (Latin & Waters, 1992).

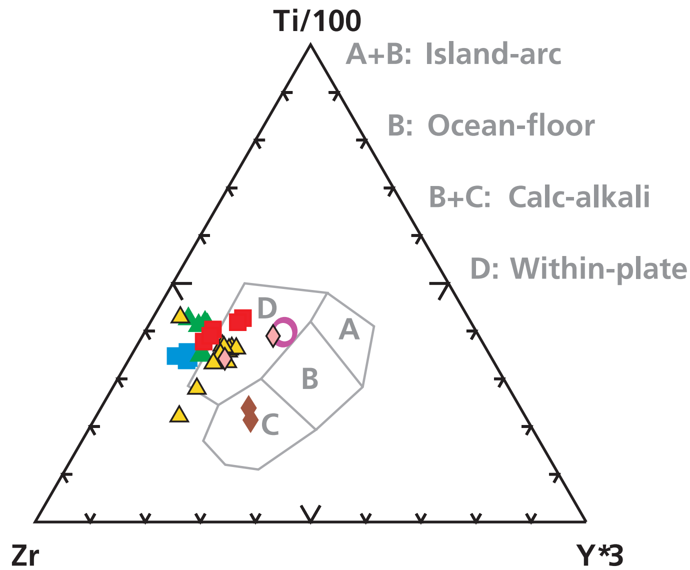

Fig. 10. Ti-Zr-Y discrimination diagram showing the tectonic affinity of mafic igneous rocks in NW Europe. Symbols as in Fig. 7; data in Table 2.

A generalised model for magma formation in the continental lithosphere can be illustrated by the example of the Zuidwal Volcano (cf. Herngreen et al., 1991). Figures 5d and e show a schematic cross section of the magma system and a hypothetical P-T diagram, which visualises the potential controls of an extension-driven melt-generation process. Given the small size of the Vlieland pull-apart basin, it is unlikely that decompression was sufficient to cause melting of a dry asthenosphere. Therefore, a more plausible scenario is that melt formed at the base of the lithosphere, which must have been enriched in volatiles and incompatible elements to lower its solidus temperature. The nature and timing of this enrichment are unknown, but may be related to a previous, possibly Permian melt-infiltration event on a regional scale. Localised stretching of the lithosphere, controlling the development of this basin, may have led to uplift of the asthenosphere-lithosphere boundary to a point where the geotherm crossed the solidus. The inferred limited degree of melting would be in line with the size of the basin. Alternatively, low-degree partial melts may have been generated in response to a regional heat pulse or uplift, their ascent to shallow crustal levels being facilitated only locally by crustal-scale faulting. The Zuidwal melt may have accumulated in a magma chamber at a relatively shallow level in the crust before and during the period when eruptions occurred, if the indications for a caldera structure (Perrot & Van der Poel, 1987) are correct.

These magmagenetic interpretations are consistent with the rift setting. The Forties basalts occur in a region that experienced the maximum lithospheric stretching at the rift triple junction. Major zones of weakness in the lithosphere may have further facilitated their rise to the surface. This would explain why there are only scattered volcanic occurrences in the adjacent Viking and North Sea Central grabens, where the degree of extension was slightly smaller. Here, asthenospheric melts may have been produced but were not able to reach the surface, and may have been underplated. An explanation for melt generation in the Netherlands and on the flanks of the North Sea Central Graben, where the amount of stretching was much smaller, can be found in the observation that in these areas the graben intersects pre-existing and long-lived stable highs such as the Texel-IJsselmeer High and the Mid North Sea High. Here, enriched continental lithosphere could have been mobilised by small amounts of asthenospheric melts to create sufficiently voluminous magma batches to reach the surface. In summary, the Jurassic-Early Cretaceous magmatic activity in the North Sea region can be attributed to rift-induced decompression melting of the upper asthenosphere and the lower lithosphere. From thermal considerations, there is no need to invoke a major mantle plume as a heat source, although a low-temperature plume cannot be ruled out (Latin et al., 1990b), particularly in the face of Mid Jurassic regional doming of the central North Sea (Ziegler, 1990).

Permian

Models for the origin of pre-Jurassic magmatism in north-west Europe are less detailed, partly because key information becomes more speculative when going further back in time. Given the absence of geochemical data on the Dutch occurrences, we will highlight only a few general points on the Early Permian, as it represents an episode of voluminous magmatism.

A conspicuous feature of the Early Permian, and locally partly Late Carboniferous, magmatism is that it started almost simultaneously in different provinces over an extensive region. Dating of volcanic and plutonic rocks from occurrences to the north of the Variscan front (Oslo Graben, Scania, the North Sea, Scotland and the North-east German Basin) shows that magmatic activity peaked in a rather narrow time interval between ca. 300 and 280 Ma, while centres in the internal Variscides as far south as Iberia and Italy are comparable in age (see Neumann et al., 2004; Timmerman, 2004; Heeremans et al., 2004, and references therein). This magmatism occurred across a collage of basement terranes with different ages, lithospheric thicknesses and geodynamic histories, and coincided with a period of wrench-related lithospheric deformation in response to a fundamental change in the regional stress field that affected western and central Europe at the end of the Variscan orogenic activity (Ziegler, 1990).

Bulk-rock data show a much greater compositional diversity as compared to that of later periods, often showing a bimodal distribution. In particular, the large abundance of intermediate and acid varieties is noticeable. Although there tends to be a regional north-south trend from strongly alkaline in the Oslo Rift to mildly alkaline in the Variscan foredeep and calcalkaline in the Rhenohercynian orogenic belt (e.g. references in Ziegler, 1990), rock types vary widely both on regional and local scales. For example, basalts encompass the entire spectrum from highly alkaline to tholeiitic in the Oslo Graben (Neumann et al., 2004), while tholeiitic as well as alkali-rich basalts have also been found in North Sea wells (Heeremans et al., 2004).

Based on trace-element and isotopic signatures, Neumann et al. (2004) suggested that the main mantle source of magmatism in the Oslo Graben, Scania and possibly the North Sea was similar to a Prevalent Mantle-type (PREMA) component residing in the lithospheric mantle. Melting was probably induced by local decompression and thinning of lithosphere in response to regional stretching north of the Variscan front, although the PREMA affinity implies that involvement of a mantle plume cannot be discarded.

Similar scenarios have been proposed for the North-east German Basin (e.g. Breitkreuz & Kennedy, 1999), where different structural domains and a heterogeneous basement added to geochemical diversity on a relatively small scale (Benek et al., 1996). Crustal thinning and block faulting facilitated the production of large volumes of intrusive and extrusive rocks. Magmas with a calcalkaline character could have been derived from a pre-existing subduction-influenced basaltic magma source (cf. Benek et al., 1996).

Locally, the thermal perturbation that was associated with the Permo-Carboniferous magmatism may have been more pronounced than in Jurassic times, as supported by the higher degrees of melting (~10%) inferred from trace-element signatures of the North Sea samples studied by Latin et al. (1990a). According to the data in Eckhardt (1979), a similar melting regime probably affected the Ems-Weser basalts (and perhaps also the Dutch Permian volcanics) in view of their Zr/Nb and Ce/Y ratios (Fig. 8). A stronger thermal anomaly is also consistent with the widespread generation of the acidic magmas, which can be explained as the products of anatexis of the lower crust, possibly provoked by the heat input from underplated basalt (Breitkreuz & Kennedy, 1999).

Cenozoic volcanism south-east of the Netherlands

Development of the Rhine rift system as part of the European Cenozoic rift system was accompanied by volcanism around the Rhine-Roer-Hessian graben triple junction in the Rhenish Massif, about 100 km south-east of the Netherlands. Volcanism started in the Eocene and lasted well into the Quaternary, with episodes of increased activity in the Late Oligocene and Miocene (Sissingh, 2003;

Dèzes & Ziegler, 2005; and references therein). It is probably the source of the volcanic material in the Lower Oligocene Boom Clay mentioned earlier. Volcanic activity and vertical movements in the area of the Rhenish Massif have been associated with the rise of a mantle plume and related thermal thinning of the mantle-lithosphere (Ritter et al., 2001).

Timing of magmatism and the opening of the Atlantic

Episodes of magmatic activity in north-west Europe, eastern North America and Greenland tend to be related to specific periods in the opening history of the Atlantic Ocean (e.g. Woodhall & Knox, 1979; Ziegler, 1988). For north-west Europe such a connection is most apparent for the post-Permian magmatic events, as (i) the rift stage of the North Sea was most pronounced between the Triassic and the Paleocene-Eocene transition, (ii) magmatism also ended by then, and had a rift-type nature throughout this time span, and (iii) the 'passive' North Sea rift system developed as a failed arm of the Artic–North Atlantic (Ziegler, 1992).

The *Middle Triassic phase* of magmatism was modest and was associated with moderate crustal stretching during the early rifting of the North Sea. It occurred before the separation of North America and Africa, which was accompanied by volcanism on the North American Atlantic coast. The *Middle Jurassic phase* of extensive volcanism at the triple junction coincided with the uplift of a major thermal rift dome in the central North Sea, possibly as a consequence of diapiric intrusion of melts at the crust-mantle boundary (Ziegler, 1992). The *Late Jurassic-Early Cretaceous phase* occurred at the peak of rifting in the entire North Atlantic region, and was roughly contemporaneous with the initial phase of crustal separation between Iberia and the Grand Banks. The volcanic activity in north-west Europe was mild, and was not accompanied by further thermal doming in the North Sea. The *latest Cretaceous-Eocene phase* is most pronounced in the north-western British Isles, Greenland, the Norwegian margin and adjacent offshore areas (e.g. Rockall), where it culminated in the Thulean volcanic event, the main phases of which lasted for some 10 Ma between late Early Paleocene and Early Eocene time, and faded away after Greenland and Europe finally separated (Ziegler, 1990). The location of Iceland near the centre of the Thulean volcanic province suggests a genetic relation between hot-spot activity and the Norwegian–Greenland Sea Rift.

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Natural and induced seismicity

B. Dost &
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ABSTRACT

Natural seismicity in the Netherlands is mainly confined to large faults in the south of the country. These faults are part of the Roer Valley Rift System, the north-western extension of the Rhine rift system. Their style of faulting, as concluded from the analysis of seismicity, is dominantly normal faulting with a small strike-slip component, and is in accordance with the current stress pattern. South of the Roer Valley Rift System the influence of large strike-slip faults and thrust faults along the Brabant Massif is felt. The effects of natural seismicity have been moderate on the whole, and include structural damage to buildings in epicentral areas. Hazard analysis based on historical seismicity shows a maximum expected magnitude of $M_L = 6.3$. Paleoseismological investigations indicate events that may reach one magnitude higher, at an average recurrence interval in the order of 2 to 3 ka along the major fault zones. In the north of the Netherlands, no significant natural activity has been detected, but since 1986 induced events are recorded, of which the locations coincide with producing gas fields. The events are shallow (1-3 km depth) and the maximum expected magnitude is estimated at $M_L = 3.8$. Despite the limited strength of the induced events, their shallow depths may result in damage to buildings, generally in small areas around the epicentres. Measured accelerations show high values, up to 3 m/s^2 at 2 km from the epicentre for an $M_L = 3.4$ event.

Keywords: Netherlands, Roer Valley Graben, active faults, induced earthquakes, source mechanisms, seismic hazard

Introduction

In the Netherlands a clear separation can be made between natural earthquakes in regions where the seismicity is caused by tectonic processes and induced earthquakes in regions where the seismicity is caused by human intervention. Natural onshore seismicity is mainly restricted to the south of the country, where faults are active. Induced earthquakes are thus far restricted to the north and are all related to the exploitation of gas fields. The mining of coal in the past and the mining of salt since the early 20th century have not resulted in noticeable seismic effects. A compilation of seismicity in the Netherlands was given by Houtgast (1991), but no comprehensive description is presently available in the literature. A regularly updated list of earthquakes in the Netherlands and immediate surroundings is maintained on internet (http://www.knmi.nl/onderzk/seismo) by the Koninklijk Nederlands Meteorologisch Instituut (KNMI, Royal Netherlands Meteorological Institute). The 1992 Roermond earthquake was covered extensively in Van Eck & Davenport (1994). Seismic events under the Netherlands continental shelf occur less frequently and relate to the Dutch Central Graben. Offshore events close to producing gas fields have been detected (KNMI earthquake list) and are suspected to be induced.

The earthquakes in the Netherlands should be viewed in a regional framework. An overview of natural seismicity for western Germany was given by Ahorner (1983), while Camelbeeck (1994) carried out a similar study for Belgium. The north-western part of Germany shows mainly events of induced origin and is of interest since these events are located by the KNMI network.

Seismological research aims to gain insight into the causes of the earthquakes, their effects at the surface and the movements that can be expected in the future. Both natural and induced earthquakes will be discussed in this chapter, each with their own characteristics and aspects of seismic hazard and risk.

Natural earthquakes

Roer Valley Rift System

Most of the natural earthquakes are located in the Roer Valley Rift System. This is an active rift system located in the Lower Rhine Embayment in the south of the Netherlands, north-eastern Belgium and western Germany (Houtgast et al., 2002). The system comprises the Roer Valley Graben and the adjacent Campine and South Limburg blocks to the south-west and the Peel and Venlo blocks to the north-east (Fig. 1). The majority of the earthquakes are confined to the Roer Valley Graben, which is bounded by two large active faults: the Peel Boundary Fault to the north-east and the Feldbiss Fault to the south-west. The main difference between these faults or fault zones is that the former is relatively narrow, up to 1 km, and the latter relatively wide, up to 5 km (Geluk et al., 1994; Houtgast & Van Balen, 2000). The earthquakes in this area are of limited magnitude, but occasionally powerful enough to cause damage. In the literature, earthquakes have been associated with known geological structures.

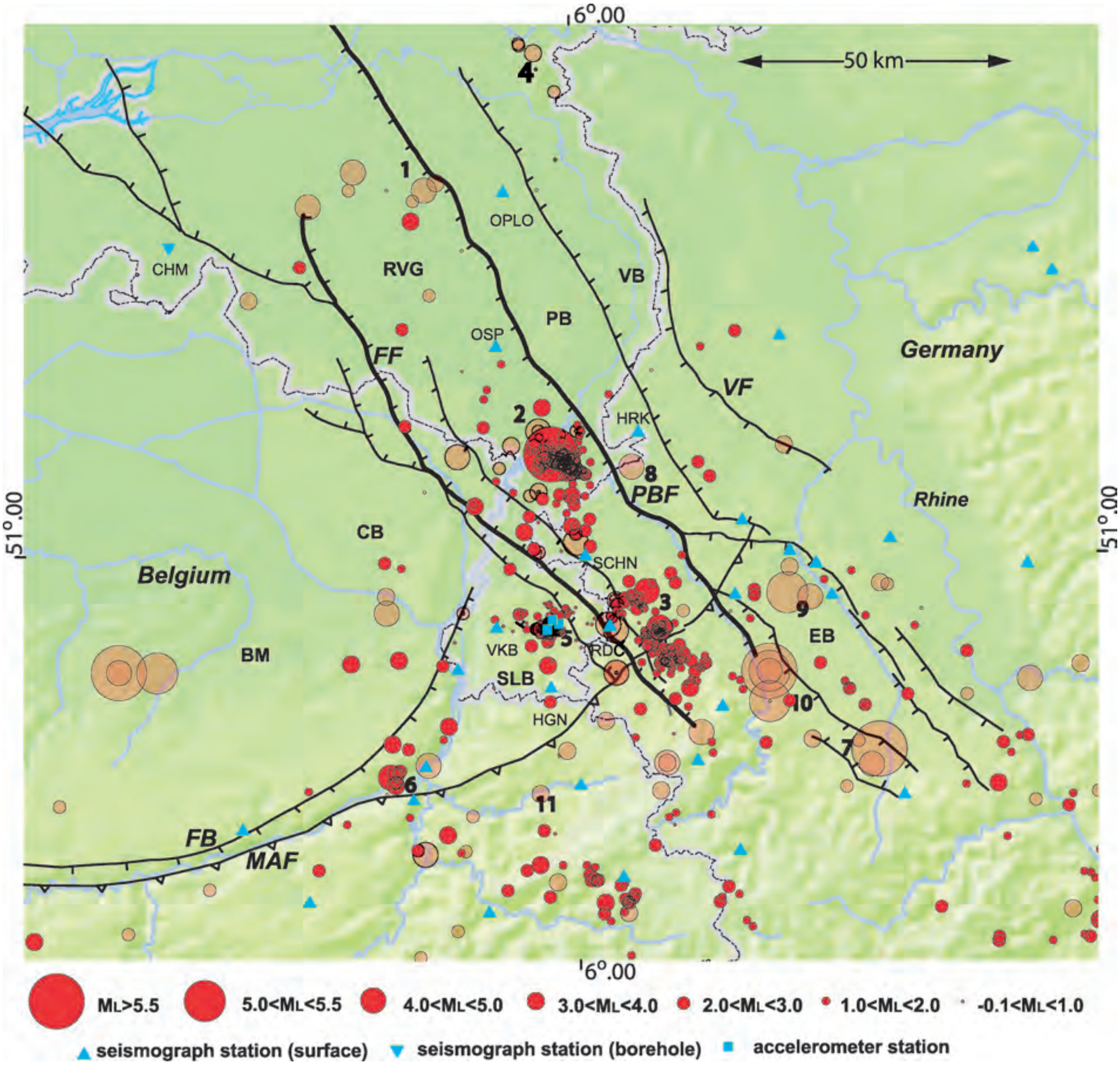

Fig. 1. Epicentre map of natural seismicity in and around the Roer Valley Graben for the period 1700-2003. Earthquakes are shown as red circles, scaled according to magnitude; events before 1980 in light red, those after 1980 in dark red. Localities mentioned in text: 1: Uden; 2: Roermond; 3: Alsdorf region; 4: Nijmegen; 5: Voerendaal; 6: Liège; 7: Euskirchen; 8: Heinsberg; 9: Tollhausen; 10: Dueren; 11: Verviers. Tectonic structures: BM: Brabant Massif; RVG: Roer Valley Graben; PB: Peel Block; VB: Venlo Block; CB: Campine Block; EB: Erft Block; SLB: South Limburg Block; PBF: Peel Boundary Fault; FF: Feldbiss Fault; VF: Viersen Fault; FB: Faille Bordière; MAF: Midi-Aachen Thrust Fault. OPLO, CHM etc. are seismic station codes.

Directly after the 1932 Uden earthquake of magnitude $M_L = 5.0$, Jongmans & Van Waterschoot van der Gracht (1932) explained this event by movement along the Peel Boundary Fault. Also, Escher (1940) mentioned the association between earthquakes and the main Roer Valley Graben boundary faults.

Peel boundary fault

The most recent, significant earthquake associated with the Peel Boundary Fault occurred April 13, 1992, had a magnitude $M_L = 5.8$, and was located just south of Roermond (Paulssen et al., 1992, Camelbeeck et al., 1994). It was felt over an area larger than 600 000 km^2 between the Czech republic, Switzerland, France and England (Haak et al., 1994), and is the strongest event in the Netherlands over the last 100 years. Its maximum observed intensity, a measure of the effects of an earthquake at the surface, was I_0 = VII (damaging). The total damage was estimated at about €128 million at 1992 value (Berz, 1994). An event of similar size occurred in 1951 near Euskirchen (Germany; Berg, 1953), 80 km SE of the Roermond epicentre. In November 1932, the second largest

Table 1. Earthquakes of magnitude $M_L \geq 4$ since 1900, related to the Peel Boundary Fault and its extension into Germany.

Event location	Date	Origin time (UTC)	Latitude (°N)	Longitude (°E)	Depth (km)	M_L
Roermond	1992-04-13	01:20:03	51.162	5.933	15	5.8
Euskirchen (D)	1951-03-14	09:46:55	50.669	6.785	28/9	5.8
Uden	1932-11-20	23:36:55	51.630	5.412	16	5.0
Alsdorf (D)	2002-07-22	05:45:05	50.875	6.213	14	4.9
Roermond	1935-01-04	04:06:13	51.140	6.145	15	4.3
Veghel	1932-11-28	05:41:35	51.610	5.640	15	4.0

Table 2. Earthquakes of magnitude $M_L \geq 4$ since 1900, related to the Feldbiss Fault and its extension into Germany.

Event location	Date	Origin time (UTC)	Latitude (°N)	Longitude (°E)	Depth (km)	M_L
Boxtel	1932-11-23	03:08:01	51.577	5.268	10	4.5
Koningsbosch/ Heinsberg (D)	1971-02-18	23:41:25	51.016	5.997	17	4.4
Stramproy	1960-06-25	14:29:15	51.158	5.687	15	4.2
Alsdorf (D)	1992-04-14	01:06:46	50.935	6.182	15	4.1

earthquake in the Netherlands, $M_L = 5.0$, occurred along the Peel Boundary Fault near Uden, 64 km NW of the Roermond epicentre (Van Dijk, 1934; Gees, 1937). There is little evidence of recent seismic activity along the fault further to the north-west. Two earthquakes, reported to be located near Tiel, ca. 90 km NNW of Roermond in 1928 and 1932 (Houtgast 1991), are based on doubtful interpretations. The 1928 event was not recorded instrumentally and was reported by only a few people in a large area. The 1932 event, recorded in De Bilt (DBN) in the Netherlands and Uccle (UCC) in Belgium, suggests a source near the Dutch-Belgian border. Therefore, the recent seismic activity along the Peel Boundary Fault appears to terminate to the north-west near Uden. The larger events since 1900 along the Peel Boundary Fault are listed in Table 1.

Aftershocks of the Roermond event enabled a detailed study of the fault plane. Camelbeeck et al. (1994) reported more than 200 aftershocks. Hypocentre locations could be calculated for 55 of them. Although the scatter in locations is high, due to a limited station coverage, an indication of the fault surface becomes clear and corresponds to an average dip of 70° to the south-west. Ahorner (1994) showed similar results in a comparable study, in which less scatter was obtained. For the Uden event one foreshock and eight aftershocks were recorded (Van Dijk, 1934). However, due to the limited instrumentation at the time, the detection threshold was much higher than for the Roermond event, and no details on the surface rupture could be obtained. Figure 1 suggests a difference in trend between the strike of the Peel Boundary Fault and the epicentre locations south of Roermond (Camelbeeck et al., 1994). This implies that the deeper structure of the fault is relatively complex. Another feature of the Roermond aftershock sequence is that two clusters of aftershocks are visible: one close to the main shock and the other 40 km to the south-east in the vicinity of Alsdorf in Germany, suggesting a triggered effect along the Feldbiss Fault (Prinz et al., 1994). The first aftershock in the distant cluster occurred only three hours after the main shock. No aftershocks are located to the north-west, where the Uden earthquake took place 60 years before. The most recent large earthquake in the Roer Valley Graben is the July 22, 2002 Alsdorf event. Its local magnitude is $M_L = 4.9$, the strongest event in the region since ten years, its depth 14 km and the location seems to correspond to the Feldbiss Fault. However, the preliminary fault-plane solution of the event (K-G. Hinzen, personal communication) shows a normal fault dipping to the south-west and suggests a connection with the Rurrand Fault, which is the extension of the Peel Boundary Fault in Germany.

Feldbiss fault

Based on the seismic activity observed in the 20th century, the Feldbiss Fault shows on average events of lower magnitude than the Peel Boundary Fault (Table 2). It is worth mentioning that a recent relocation of the Koningsbosch event shifts its location into Germany, close to Heinsberg, which is in accordance with the findings of Ahorner & Pelzing (1983). For some events that are located in the middle of the Roer Valley Graben it is not clear whether they belong to the Peel Boundary Fault or the Feldbiss Fault.

Table 3. Deep earthquakes since 1900 south of the Feldbiss Fault ($M_L > 2.0$).

Event location	Date	Origin time (UTC)	Latitude (°N)	Longitude (°E)	Depth (km)	M_L
Gulpen	1988-10-17	19:39:54	50.813	5.922	23	3.5
Verviers (B)	1972-02-17	04:03:32	50.671	5.970	22	3.1
Valkenburg	1981-12-20	10:38:53	50.852	5.922	23	2.7
Sippenaeken	1995-07-18	12:51:23	50.752	5.927	23	2.4
Valkenburg	1995-03-30	20:29:08	50.846	5.878	20	2.2
Maastricht	1990-10-18	19:30:29	50.813	5.647	21	2.2

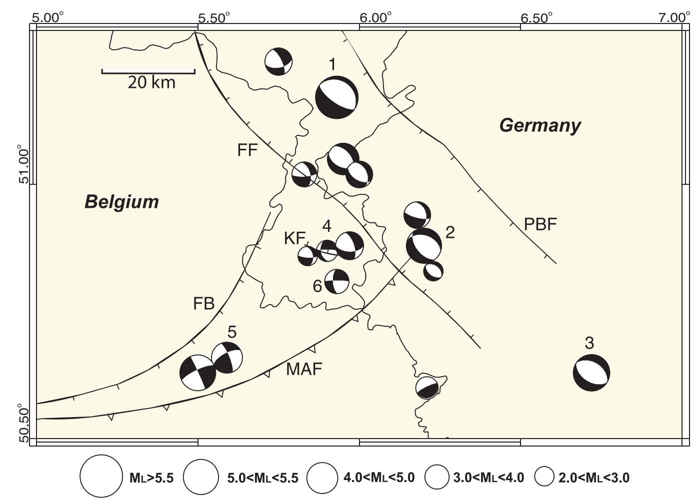

Fig. 2. Overview of source mechanisms in and around the Roer Valley Graben, represented as equal-area projection of the lower focal hemisphere. Quadrants of compressional P-wave onsets are filled. Abbreviations of fault names as in Fig. 1; KF: Kunrade Fault. Localities of events: 1: Roermond; 2: Alsdorf; 3: Euskirchen; 4: Voerendaal; 5: Liège; 6: Gulpen.

AREA SOUTH OF THE FELDBISS FAULT

The seismicity south of the Roer Valley Graben is spread over a relatively wide area. The largest event in the 20th century is that of magnitude $M_L = 5.0$ on November 8, 1983 at Liège in Belgium (Camelbeeck & De Becker, 1985). This shallow event, only 6 km deep, is connected to the eastern prolongation of the Brabant strike-slip zone (Ahorner, 1985), between the Midi-Aachen Thrust Fault and the Faille Bordière. The South Limburg Block is characterized by events that originate from the lower crust at depths of 20 to 24 km (Table 3; KNMI earthquake list).

The region around Voerendaal, just south of the Roer Valley Graben, shows an anomalously shallow, swarm-like seismicity at depths between 3 and 8 km. A series of events were recorded during more than one month, starting December 7, 1985, the largest of them having a magnitude of $M_L = 3.0$. After a period of 15 years, a new swarm of shallow events started in the same area on December 20, 2000. This swarm lasted until January 2002 and a total of 139 events with magnitudes ranging between $M_L = 0.1$ and 3.9 were recorded by a seismograph network that had been significantly improved with respect to detection capability. The largest event of this swarm occurred during the middle of its activity. The activity is most probably connected to the E-W oriented Kunrade Fault (Fig. 2), which shows a scarp at the surface just south of Voerendaal. The shallow occurrence of the events explains the relatively high maximum intensities, ranging from $I_0 = III$ (weak) for $M_L = 2.0$, to $I_0 = VI$ (slightly damaging) for $M_L = 3.9$. The strongest event caused some damage to structures, like broken or fallen chimneys and roof tiles. This kind of damage is comparable to that caused by the much stronger Uden and Roermond events, which occurred at larger depths. The combination of magnitude and depth largely determines the intensity at the surface.

Other areas

Instrumentally recorded natural seismicity onshore outside the region discussed above, is limited to a few locations near Nijmegen, ca. 70 km north of Roermond, where two events of magnitude $M_L = 3.2$ and 3.0 have been recorded in 1972 and 1979. It is unclear whether they can be associated with the Viersen Fault of the Roer Valley Rift System. In the area south of Bergen op Zoom, close to the Belgian-Dutch border, a small event of magnitude $M_L = 2.1$, possibly connected to the Antwerpen Fault, took place on June 15, 1988.

The North Sea area is subject to large events, as demonstrated by the 1931 Dogger Bank event, $M_L = 6.0$, that was felt all over the Netherlands (Gees, 1937). An overview of the seismicity in the area is given by Ringdal (1983). Recent seismicity in the Netherlands continental shelf is limited to on average one event per year along existing faults. For this region the lowest-magnitude event that can be detected is estimated at $M_L = 3.0$.

Instrumentation and location accuracy

The KNMI started seismological observations in De Bilt (station code DBN) near Utrecht, in 1904 on an experimental basis (Houtgast et al., 1987). In 1908 it started continuous recording, and a historical archive of analogue seismograms is kept and maintained since then.

In 1926 a second station was installed in Heerlen (HEE, E-W component only), in order to investigate the local seismicity in the south of the country. Instruments used at that time were heavy and not very portable. In 1951 a third station was added near Witteveen (WIT) in the north (Drenthe), followed since 1970 by additional stations in the south: Ravensbosch (RSB), Epen (ENN), Kerkrade (KRN) and Valkenburg (VKB). All data of these stations are recorded on paper. In the 1970s, experiments were carried out with digital recording and a modernization of the network was planned. In 1993, station Heimansgroeve (HGN) near Epen started its operation after a careful site selection and became the new low-noise reference station for the Netherlands. The station is equipped with state-of-the-art, very broad-band sensors and digital recording (Dost & Haak, 2002). At present three short-period digital stations are operational in southern Limburg in addition to HGN: Valkenburg (VKB), Rolduc (RDC) and Schinveld (SCHN). The northern part of Limburg is monitored by two broad-band stations: Oploo (OPLO) and Herkenbosch (HRK). Finally, station Winterswijk (WTSB), an upgrade of the 1974 station WTS in east Gelderland, completes the picture to the north. Apart from the KNMI instrumentation, a set of mobile stations, the NARS array, has been operated by the University of Utrecht in variable configuration since 1983 (Dost et al., 1984). In the period 1989-1993 the NARS stations have been deployed in the Netherlands, Belgium and Germany. In Belgium the station density increased after the Liège earthquake of 1983, while at the same time also in Germany more stations were added by the University of Köln and the Geological Survey of Nordrhein-Westfalen.

Before the 1930s, the epicentre locations of most earthquakes were based on seismic intensity data, due to the low density of seismograph stations at the time. Errors in these locations can easily be in the order of several kilometres, especially if the epicentral area was thinly populated. Assessment of the depth of the events caused problems until the 1980s, when the number of seismological instruments increased. A good example is the Euskirchen 1951 event, of which Berg (1953) calculated a depth of 28 km based on instrumental data, while Sponheuer (1958) and Van Gils & Zaczek (1978) calculated a shallow depth around 9 km based on intensity data. Since the 1980s the average error in epicentre location is ca. 1 km, while the depth accuracy is in the order of several kilometres. In some instances of a good correlation between waveforms of different events at the same station, an increased resolution can be obtained, improving the relative epicentre location to 0.1 km. Until now such improved resolutions have only been obtained for induced earthquakes near two gas fields (Bergermeer and Roswinkel; see section on 'Induced earthquakes', which also mentions special instrumentation).

Although the station density increased, routine data analysis is still based on an average crustal seismic velocity model. Until recently, the velocity model of Ahorner (1983) was used to locate earthquakes in the southern Netherlands. Reamer & Hinzen (2004) derived a new one-dimensional model, based on a digital version of the phase database from the Bensberg network in western Germany. A joint inversion technique using a combined phase database from Bensberg, De Bilt and Uccle may further refine the velocity model(s) and improve the location accuracy of events in the region.

Magnitude

Magnitude is a measure of the strength of an earthquake at its source, while intensity describes the effects of an earthquake at various locations at the surface. Intensity varies with epicentral distance, so the maximum intensity is used to characterize an earthquake. Richter's (1935) definition of the local magnitude (M_L) was used in different ways by different agencies and has not always been applied to older data. Instead, magnitudes were often estimated from reported maximum intensities, especially for older events, and may therefore vary up to one magnitude unit. In his overview of earthquakes in the Netherlands, Visser (1943) gave no magnitudes, only maximum intensities. Houtgast (1991) estimated magnitudes for events prior to 1900 from the maximum intensity by simple relations originally derived for southern California (Båth, 1979, p. 127). De Crook (1994) derived and used an empirical relation between intensity and local magnitude M_L for the Netherlands, based on 42 observed tectonic earthquakes.

Hinzen & Oemisch (2001) applied a more sophisticated method to estimate location and magnitude directly from the raw seismic intensity data and used recent events (including Roermond 1992) as a training set to calibrate the method. This method provides an objective and reproducible way to interpret data and was used to re-evaluate older events in the northern Rhine area. Results for pre-1900 events will be evaluated in future updates of the catalogue of events in the Netherlands.

Richter's local-magnitude M_L scale is based on earthquake recordings from a standard short-period horizontal seismograph (Wood-Anderson). The M_L scale saturates for events larger than magnitude 6.5 (McCalpin, 1996) since the dominant energy of the recorded seismic waves is lower than the pass band of the seismograph. An alternative is to use the moment magnitude M_w (Hanks & Kanamori, 1979), which is a measure based on the seismic moment, defined as the product of the average displacement over the fault zone, the area of the fault surface and the average shear rigidity of the faulted rocks. M_w is derived from the frequency spectra of seismograms and does not saturate. For the 1992 Roermond earthquake, where the local magnitude M_L is expected to be close to the moment magnitude M_w since saturation is not yet important, it appears that the local magnitude (5.8) is much larger than the moment magnitude (5.4). This incompatibility is not yet solved, but may cause confusion between reported magnitudes of events. Reamer & Hinzen (2004) made a recent comparison between M_w and M_L based on broad-band seismograms for events from the past ten years. Their paper supports the statement that $M_L > M_w$ for events in the Lower Rhine Embayment, with a tendency for the difference $M_L - M_w$ to increase for larger events.

Source mechanisms

Apart from the locations of earthquakes, other source parameters can also be derived from recorded seismograms. The derived source mechanism, i.e. the direction of movement along the fault plane, combined with an accurate location, allows to relate events to faults and to gain insight into the source processes. An overview of determined source mechanisms in and around the southern Netherlands is shown in Figure 2.

The Roer Valley Graben is dominated by normal faulting; the graben subsides with respect to its flanks. The hypocentral depths of earthquakes in the graben range from 5 to 20 km, and indicate that the boundary faults continue into the lower crust. In the South Limburg Block, events occur mainly at depths between 20 and 24 km (Table 3), and the mechanism appears to change from predominant normal faulting to strike-slip faulting (Camelbeeck, 1994). However, since these deep events are usually of low magnitude, their source mechanisms are not well constrained. This possible change in mechanism would imply a different cause for these earthquakes and may be related to large strike-slip faults in Belgium (Camelbeeck & Van Eck, 1994) or to differences in stress regime between upper and lower crust. SW-NE striking thrust faults, like the Midi-Aachen Fault south of the Brabant Massif (Hance et al., 1999; Mansy et al., 1999), show their influence in north-eastern Belgium and western Germany (Ahorner, 1983) through earthquakes with well-constrained thrust mechanisms.

For the Voerendaal events, different source solutions have been made. For the largest 1985 event, Houtgast (1991) reports a normal fault of 270° strike and 80° N dip. Camelbeeck (1994), however, reports a strike-slip movement. A similar controversy applies to solutions for the largest shocks in the 2000-2002 swarm, although the normal-faulting solution is best constrained and here accepted as typical for these events. The Voerendaal events are connected to the Kunrade Fault, that deviates from the general NW-SE trends in the Roer Valley Graben and shows an E-W strike.

Plenefisch & Bonjer (1997) carried out a stress inversion for the entire Rhine Graben based on 98 focal mechanisms of small to moderate events. Although no definite distinction could be made between strike-slip and normal faulting regimes, a counter-clockwise rotation of the horizontal stress axes of 30° could be observed between the southern Rhine Graben (160°) and the Roer Valley Graben (130°). A separation of the total dataset for the southern Rhine Graben in an upper-crust (5-15 km) and a lower-crust (15-23 km) part showed a dominant strike-slip regime in the upper crust, and a normal faulting regime in the lower crust. A similar study for the northern Rhine Graben (Hinzen, 2003) shows the opposite: a dominant normal faulting regime in the shallow crust (< 12 km) and a strike-slip regime in the deep crust.

Seismotectonic setting

Ziegler (1992, 1994) describes the evolution of the Roer Valley Rift System as part of a Cenozoic European rift system. Two factors control the seismotectonic movements in the rift: first the present NW-SE compressive regime, characterized by an orientation of the maximum horizontal stress of 145°, and controlled by push from the Mid-Atlantic Ridge and by the collision of Africa and Europe (Gruenthal & Stromeyer, 1992; Goes et al., 2000). A second factor is the existence of major fault zones (Ahorner, 1994) of post-Hercynian or even older origin that have been reactivated during later tectonic phases (De Jager, this volume). Mueller et al. (1997) discuss the observation that source mechanisms vary in western Europe over short distances from strike-slip to normal faulting, while the general strike of the faults does not change. This is hard to understand if the regional stress field remains the same and can only be explained by a decoupling of the brittle upper crust from the lithospheric mantle by a ductile lower crust. It also explains the above-mentioned different styles of faulting in the upper and lower crust interpreted by Plenefisch & Bonjer (1997). The occurrence of deep and shallow events south of the Feldbiss zone may also be explained by this decoupling, although an alternative explanation may be the existence of a deeper set of faults at a different strike. Unfortunately, we know little about the deeper geology of south Limburg, and only the largest deep event, Gulpen 1988, has a well-determined source solution (strike-slip). If decoupling is assumed, the model needs to explain why normal faulting is dominant in the upper crust and strike-slip in the lower crust, just opposite to the findings in the southern Rhine Graben. Hinzen (2003) explains the difference by interpreting the shallow extensional regime as a local effect superimposed on the regional strike-slip regime. Camelbeeck & Van Eck (1994) argued that the deep Gulpen event and the shallow 1983 Liège event may belong to the same tectonic region, which may be an alternative explanation.

The Roer Valley Rift System has been studied in detail with respect to its recent and long-term, subsidence and uplift history (Zijerveld et al., 1992; Van den Berg et al., 1994; Geluk et al., 1994; Houtgast & Van Balen, 2000) and to rheological and geomechanical modelling (Dirkzwager et al., 2000). Subsidence rates that differ over time may be related to changes in the seismicity. In the Netherlands, subsidence rates are based on geodetic and geological measurements. The subsidence rate in the Roer Valley Graben was not constant; periods of acceleration and deceleration occurred since the last 8 Ma (Geluk et al., 1994). This has to be taken into account in the interpretation of paleo-seismicity (e.g. Houtgast, 2003). The geomechanical modelling showed that the boundary faults can be reactivated at reduced friction angles. Details of this modelling are given by Dirkzwager (2002), who modelled the subsidence in and around the graben by such reactivation. Rheological models of a layered crust were developed by Cloetingh & Burov (1996) and used by Mayer et al. (1997) to explain the difference in behaviour between upper crust (strike-slip) and lower crust (normal faulting) for the southern Rhine Graben.

In general terms the earthquakes can be associated with the major fault zones. However, little is known about the extent of the faults at depth. Ahorner (1983) combined seismicity data with seismic reflection data and showed that shallow (5 km deep) thrust faults are active to the south-east of station HGN. Later information from deep seismic reflection studies in Germany and Belgium (DEKORP Research Group, 1991) confirmed his interpretations. In the Netherlands, only limited information is available from deep seismic surveys (Duin et al., 1995). They provide little information on the deep structure of southern Limburg. For most of the region south of the Feldbiss Fault, the structure below 200-300 m is largely unknown (NITG, 1999). This means that the association between seismicity and geology in the region around Voerendaal cannot yet be established.

Historical and paleo-earthquakes

The evidence for the occurrence of historical earthquakes comes mainly from written sources. For the 19th century these sources are in general clear, but for earlier centuries the help of historians to interpret them is required (Alexandre, 1985, 1994), and the accuracy of reconstructed epicentre locations (10-20 km) and magnitudes (1-2 units) becomes limited. A few 19th-century events are worth mentioning (Fig. 1). On September 26, 1878, an event occurred near Tollhausen in Germany, with an estimated magnitude $M_L = 5.6$. This event is possibly connected to the northern boundary fault of the Erft Block (Hinzen & Oemisch, 2001). One year before, on June 24, 1877, an event of magnitude $M_L = 4.4$ occurred near Herzogenrath, where it was preceded in 1873 by a magnitude $M_L = 4.3$ event. Both were most probably connected to the Feldbiss Fault. In the 18th century a large earthquake of magnitude $M_L = 6.4$ happened near Düren (February 18, 1756), and at the end of the 17th century a major event, possibly of magnitude $M_L = 6.8$, occurred near Verviers in Belgium (September 18, 1692; Hinzen & Oemisch, 2001).

At most, we have a reasonable impression of the larger seismic events over the last 500 years. The question arises: did we miss any of the large events in the Netherlands that are expected to occur on average once every 2000 or 3000 years? Paleoseismic research is directed at solving this question. A first survey for paleo-earthquakes along the Peel Boundary Fault has been carried out recently (Demanet et al., 2001; Van den Berg et al., 2002). The survey aimed to collect evidence for sudden displacements along the fault zone, where it comes to the surface. These displacements can be found in trenches of a few metres deep, oriented perpendicular to the fault plane. Since the preparation of a trench is time-consuming, geophysical methods were applied to locate the fault at the surface. Near Neer, north-west of Roermond, a trench was prepared intersecting the Peel Boundary Fault. It showed geological evidence for two, or possibly three paleo-earthquakes. The strongest events are dated ca. 15 ka BP with a time interval between them of ca. 2 ka. Their moment magnitudes are estimated at $M_w = 6.2$ to 6.5, which is a magnitude larger than the 1992 Roermond event. Similar results are reported from surveys in Belgium along the Feldbiss Fault (Camelbeeck & Meghraoui, 1998).

Natural seismic hazard and risk

Based on a catalogue of earthquakes in the region, De Crook (1993) carried out a probabilistic hazard assessment for the Netherlands in terms of the probability of intensity recurrence. Important model parameters in this assessment are the definition of the seismotectonic zonation, the relation between maximum observed intensity and number of events, and the intensity–distance attenuation relation. The results of the assessment are presented in a series of maps showing the probability of occurrence per year for the expected intensity. The 1992 Roermond earthquake corroborated the assessment nicely (De Crook, 1994). However, engineering models are better in handling ground-acceleration estimates than intensities. De Crook (1996) presented first rough estimates of expected horizontal peak ground accelerations, based on empirical intensity–peak ground acceleration relations used in Germany.

The statistical relation between the logarithm of the cumulative annual number of events and magnitude, usually referred to as the frequency–magnitude relation, shows for the southern Netherlands a characteristic linear part for magnitudes between $M_L = 2.5$ and 4.5. The slope of the regression line, called the b-value, is used to estimate mean return periods for events and may be used to estimate the maximum expected magnitude. De Crook

(1996) determined a b-value of 0.67 ± 0.03 for the seismicity in the south of the Netherlands, which is comparable to the b-value of 0.74 determined by Ahorner (1983) for the Rhenish Massif, implying that the seismicity in the south of the Netherlands is comparable to the seismicity in the larger region. De Crook further estimated the maximum expected magnitude for the south of the Netherlands at $M_L = 6.3$.

The use of a dense seismometer network allows to calculate locations, magnitudes and source mechanisms. The reconstruction of true ground motion in the epicentral area, however, is less accurate since attenuation and site effects on a local scale are not well known. This realization motivated KNMI to install two accelerometers in the epicentral area after the first earthquake near Voerendaal (December 2000), where a new earthquake swarm could be expected. Accelerometers are capable of recording strong motions, where seismometers are saturated. Later, two more accelerometers were added in order to study local variations in ground movement. Acceleration data are important for the determination of seismic hazard and for estimation of the maximum expected ground movement. A new scaling relation of peak ground acceleration and peak ground velocity for the Netherlands has recently been constructed (Dost et al., 2004). In addition the accelerations are used in modelling studies to determine the vulnerability of houses and infrastructure. This relates to earthquake risk, which is equal to the product of hazard and vulnerability. Vulnerability is defined as the probability of damage as a result of ground movement. In times when economic and insured losses from earthquakes increase dramatically (Allman & Smolka, 2001), risk evaluation is gaining importance.

Induced earthquakes

History, instrumentation and examples

The northern provinces of the Netherlands are considered to be aseismic (Houtgast, 1991; Haak, 1993). However, since 1986 seismic activity has been recorded in and around producing gas reservoirs in these provinces. Up until December 2002, over 280 weak earthquakes were recorded (Fig. 3). Twenty-six of these were larger than $M_L = 2.5$ (Table 4), the detection threshold for the region prior to 1991. The first event, at Assen in December 1986, with $M_L = 2.8$, was followed nearly one year later by a tremor a few kilometres south of Hooghalen ($M_L = 2.5$; Fig. 3). At the time, it was not clear what caused these earthquakes and a connection with the production of natural gas was suggested. Similar relations between earthquakes and producing gas fields were noted in France (Grasso & Wittlinger, 1990) and Canada (Wetmiller, 1986). The Ministry of Economic Affairs, responsible for the granting of concessions to the gas-producing companies, asked KNMI to monitor the region. A small-scale seismic network covering an area of 400 km^2 was installed around Assen in 1988. The network consisted of six vertical-component seismic stations, located at the surface. At the end of 1989, a third tremor ($M_L = 2.7$) occurred in Noord-Holland near Purmerend, some 100 km to the west. The problem appeared to be more spread out and the need for more effective monitoring was felt.

Noise conditions in the northern Netherlands do not allow the detection of small seismic events. Within the Assen array the detection threshold was estimated at magnitude $M_L = 1.7$. In order to obtain a similar or even improved threshold over a larger area, borehole seismometers were installed in 1995, after extensive testing in a site near Finsterwolde (FSW), east of the Groningen gas field. In boreholes the ground noise diminishes rapidly with depth (Carter et al., 1991). A reduction in noise by a factor of almost ten in the frequency band between 1 and 10 Hz was achieved in Finsterwolde, showing for the first time dozens of minor tremors in the Groningen field (Haak, 1993). In total, eleven borehole seismometers, eight in the provinces of Groningen and Drenthe and three around Alkmaar in Noord-Holland, were installed. The boreholes are 200 m deep, have no casing, and contain a minimum of four levels of 3-component, low-frequency, 4.5-Hz geophones, electronically modified to a 1-Hz corner frequency. The total network shows a detection threshold of magnitude $M_L = 1.5$, although much smaller tremors have been detected close to some stations.

The induced earthquakes in Groningen and Drenthe are all located in or near producing gas fields (Fig. 3). The average error in location in the horizontal plane is ca. 1 km. For the Groningen field the locations cluster around large faults in or at the edge of the reservoir which produces from a depth of ca. 3 km. Remarkably, the south-eastern part of the field shows no seismic activity at all. Other gas fields with significant seismic activity are Annerveen, south of the Groningen field, several NNW-SSE trending smaller fields to the west, and the Emmen and Roswinkel fields in south-east Drenthe.

The activity around Roswinkel gained special attention. In the period 1992-2002, a total of 35 earthquakes occurred with magnitudes between $M_L = 0.8$ and 3.4. Events of magnitude 2.0 and higher are felt by people in the village, and minor damage was reported for the events of largest magnitude. Since the seismograms of these events have a high coherency, the locations could be obtained at a high resolution in the horizontal plane, in the order of 0.1 km or less, but the depth remains uncertain. Figure 4 shows the locations of the Roswinkel events, the gas field and the faults at the top of the reservoir. The events clearly follow both the WSW-ENE, and the NW-SE striking fault system. Both systems are explained by wrench faulting (Frikken, 1999). The source mechanisms for these events

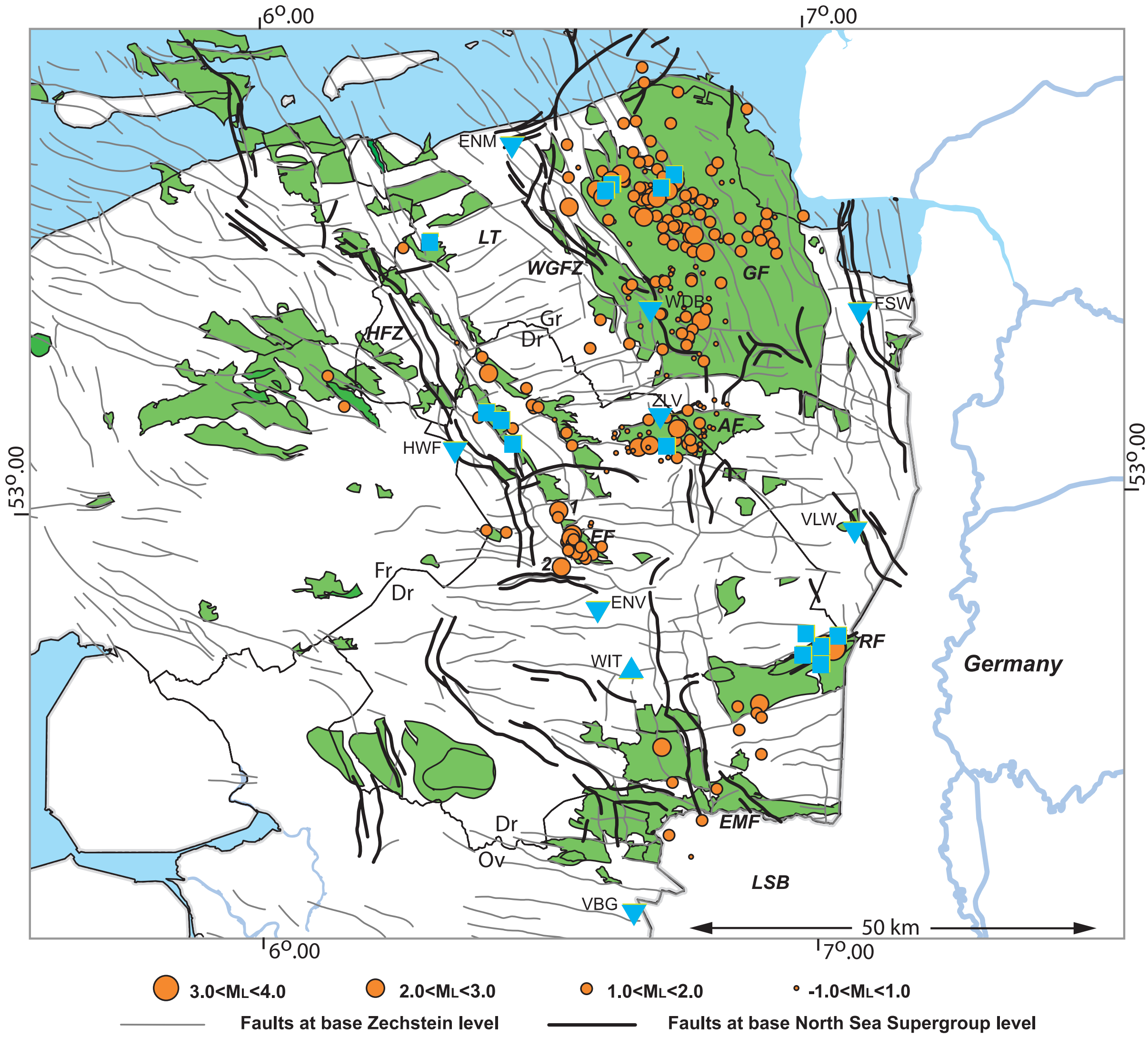

Fig. 3. Epicentre map of induced earthquakes in the north-eastern provinces of the Netherlands: Friesland (Fr), Groningen (Gr), Drenthe (Dr) and Overijssel (Ov). The first two events are marked: 1: Assen (1986), 2: Hooghalen (1987). Gas fields are shown in green: GF: Groningen field; AF: Annerveen; RF: Roswinkel; EF: Eleveld; EMF: Emmen. Faults shown are at base Zechstein, corresponding to the average hypocentral depth of ca. 3 km, and at base North Sea Supergroup (Tertiary) at ca. 0.7 km depth. Blue symbols indicate seismic stations (with their station codes) and accelerometer stations (legend in Fig. 1). Major tectonic structures: HFZ: Hantum Fault Zone; WGFZ: West Groningen Fault Zone; LT: Lauwerszee Trough; LSB: Lower Saxony Basin.

are assumed to be very similar since the waveforms are highly coherent. There are two possibilities: the preferred solution from a seismological point of view is a low-angle thrust, while a steep reverse fault may be an alternative. The latter case would correspond to reverse reactivation of a normal fault at the top of the reservoir. In reflection seismic profiles vertical faults and shallow low-angle thrust faults are both visible.

To improve depth resolution of seismic events one could add stations in the area, or deploy sensors in well bores close to the reservoir. In 2001 the Nederlandse Aardolie Maatschappij (NAM) initiated a 'downhole microseismic' experiment near Roswinkel, with its reservoir at a depth of approximately 2 km. A total of 29 minor events with magnitudes between $M_L = -6$ and -2 could be located. They confirmed a depth of 2 km and in map view they follow the same pattern as the larger events detected by the KNMI shallow-borehole and surface stations (Fig. 4). A later downhole experiment in the Annerveen field was less successful due to technical problems. Nevertheless, ten events were recorded of magnitudes less than $M_L = 0.9$.

Table 4. Larger ($M_L \geq 2.5$) earthquakes in the north of the Netherlands.

Nr	Location	Date	Origin time (UTC)	Latitude (°N)	Longitude (°E)	M_L	Depth (km)	I_0
1	Assen	1986-12-26	07:47:51.0	52.992	6.548	2.8	1	IV-V
2	Hooghalen	1987-12-14	20:49:51.0	52.928	6.552	2.5	1.5	IV
3	Purmerend	1989-12-01	20:09:14.3	52.529	4.971	2.7	1.2	V
4	Geelbroek	1991-04-25	10:26:31.5	52.952	6.575	2.6	3	III-IV
5	Eleveld	1991-08-08	04:01:14.6	52.965	6.573	2.7	3	III-IV
6	Geelbroek	1992-05-23	15:29:11.4	52.953	6.572	2.6	3	III-IV
7	Roswinkel	1992-06-11	17:09:37.0	52.831	7.032	2.7	2	III-IV
8	Eleveld	1992-07-22	23:23:13.2	52.961	6.570	2.5	3	III
9	Roswinkel	1994-02-05	15:10:05.8	52.833	7.045	2.9	2	IV-V
10	Stedum	1994-07-01	06:27:42.6	53.332	6.577	2.7	3	I
11	Middelstum	1994-07-30	09:18:20.7	53.351	6.628	2.7	0.9	IV-V
12	Alkmaar	1994-08-06	18:02:19.2	52.654	4.711	3.0	3.0	IV-V
13	Alkmaar	1994-09-21	01:12:58.1	52.658	4.708	3.2	3.2	V
14	Roswinkel	1995-06-20	08:59:40.1	52.832	7.029	2.7	2	III
15	Roswinkel	1996-03-12	12:13:48.7	52.838	7.059	2.6	2	IV
16	Roswinkel	1996-12-28	18:16:52.7	52.834	7.043	2.7	2	IV-V
17	Roswinkel	1997-02-19	21:53:50.8	52.832	7.038	3.4	2	VI-VII
18	Roswinkel	1998-01-28	21:33:03.8	52.833	7.040	2.7	2	V
19	't Zandt	1998-02-15	07:24:16.4	53.356	6.773	2.6	3	III
20	Roswinkel	1998-07-14	12:12:02.2	52.833	7.053	3.3	2	V
21	Roswinkel	1999-12-31	11:00:55.3	52.835	7.048	2.8	2	V
22	Loppersum	2000-06-12	15:48:23.0	53.340	6.742	2.5	3	III
23	Roswinkel	2000-10-25	18:10:34.7	52.832	7.052	3.2	2	V
24	Alkmaar	2001-09-09	06:58:12.6	52.651	4.713	3.5	2	VI
25	Alkmaar	2001-09-10	04:30:15.3	52.653	4.712	3.2	2	V
26	Bergen aan Zee	2001-10-10	06:41:09.3	52.682	4.648	2.7	2.9	III

I_0 is the intensity in the epicentre.

In Noord-Holland, between Alkmaar and Bergen, two earthquakes of magnitude $M_L = 3.0$ and 3.2 happened in 1994 (KNMI, 1994a, b). They were located along a NW-SE trending fault in the Bergermeer gas field (Roest & Mulders, 2000). In 2001, after seven years of quiescence, the area was confronted with two more events of magnitude $M_L = 3.2$ and 3.5 (Fig. 5). Analysis showed that the four events are located at a fault that runs from the reservoir to the south-east (Haak et al., 2001). This fault is a normal fault, which is reactivated in reverse direction by compaction of the gas-producing reservoir in the upper block. The source mechanism is well determined, as is explained further on. Also in 2001, a first event was recorded at the edge of the Bergen gas field, west of Bergermeer.

The KNMI borehole network shown in Figure 3 also records events induced in north-western Germany. Events of magnitudes between $M_L = 2.5$ and 3.0 are located near Cloppenburg, where gas fields are in exploitation. During 2002, the area south of Bremen also showed seismic activity related to gas production (Leydecker, 2002). The area around Ibbenbueren is active since 1980. The events are shallow (1-2 km), their reported magnitudes reach up to $M_L = 4.3$, and they are located in a coal-mining area.

Seismotectonic framework

The induced earthquakes seem to be correlated with faults within or at the edges of producing gas fields. A reactivation of existing zones of weakness is, like in the case of natural earthquakes, a preferred mechanism. Therefore, the tectonic framework is briefly described below.

In the north-eastern Netherlands the tectonic structure is dominated by the Lauwerszee Trough, bounded by the NW-SE trending Hantum and West Groningen fault zones (RGD, 1995; Frikken, 1999; De Jager, this volume). Figure 3 shows that the large Groningen field is situated east of the Lauwerszee Trough on the Groningen Block, while the smaller fields to the west are more or less close to the Hantum Fault Zone. Within the trough a series of WSW-ENE trending faults show dominantly strike-slip movements (Frikken, 1999). Only one recent (17-5-2001) event has been located near one of these strike-slip faults. It is too small and data coverage too insufficient to determine a source mechanism. Only a few events are located west of the Hantum Fault Zone. South of the Eleveld gas field the tectonic structure becomes dominated by the E-W trending faults of the Lower Saxony Basin (De Jager, this volume).

The region around Alkmaar belongs structurally to the Central Netherlands Basin. The published geological maps (RGD, 1993) show faults at reservoir level, but their accuracy is limited. New maps, based on 3D-seismic surveys, and made available by BP-AMOCO, show a clear connection between seismicity and faults at reservoir level (Fig. 5).

Mechanisms of induced events

In October 1991, a multidisciplinary research programme of two years was started to clarify the cause of the earthquakes that appeared to be associated with gas fields. The final report concluded that the earthquakes could have been induced by gas production (Haak, 1993). This conclusion was based on the seismic pattern, i.e. hypocentral locations, which was not consistent with a purely natural tectonic origin, on the time sequence of the events, and on the pressure histories of the reservoirs.

The relationship between gas production and earthquakes was further investigated, using the knowledge that existing faults form zones of weakness in a changing stress field. The analysis was done by calculating with a numerical model the stress changes resulting from pressure drops in gas reservoirs (Roest & Kuilman, 1994). Its results showed that there are three mechanisms by which gas extraction can reactivate faults. Most events can probably be explained by the mechanism that relates to faults located either within gas reservoirs or at their margins. Gas, which is accumulated in porous rock at a depth of approximately 3 km, is under a local pressure of about 350 bar (35 MPa). This is a sizeable fraction of the total pressure of ca. 750 bar within the gas reservoir. In the northern Netherlands, the total pressure within the Earth's crust is almost completely due to the weight of the overlying rock (Rondeel & Everaars, 1993). The horizontal stress component, which results from the large-scale movement of the tectonic plates, is limited because of slow plate motion. The stress regime is extensional in a SW-NE direction, similar to that in the southern Netherlands.

During production, the gas pressure in the reservoir (ca. 350 bar at 3 km depth) gradually declines. The pressure exerted by the overburden (ca. 750 bar) remains the same. As a result, the reservoir undergoes compaction (Geertsma, 1973; Geertsma & Van Opstal, 1973). This compaction at depth results in ground subsidence at the surface, which is widely observed and causes concern for the water management in the area. Whereas ground subsidence tends to equal the compaction at reservoir depth in the central areas of large gas fields, it is only a small fraction of such compaction in small fields. Large and small in this respect refer to the average radius of the gas field in relation to the reservoir depth.

Earthquakes can occur where the compaction on one side of the fault differs from that on the other. This differential compaction may occur at the edge of a gas reservoir where it is sealed against a fault, or at faults within the

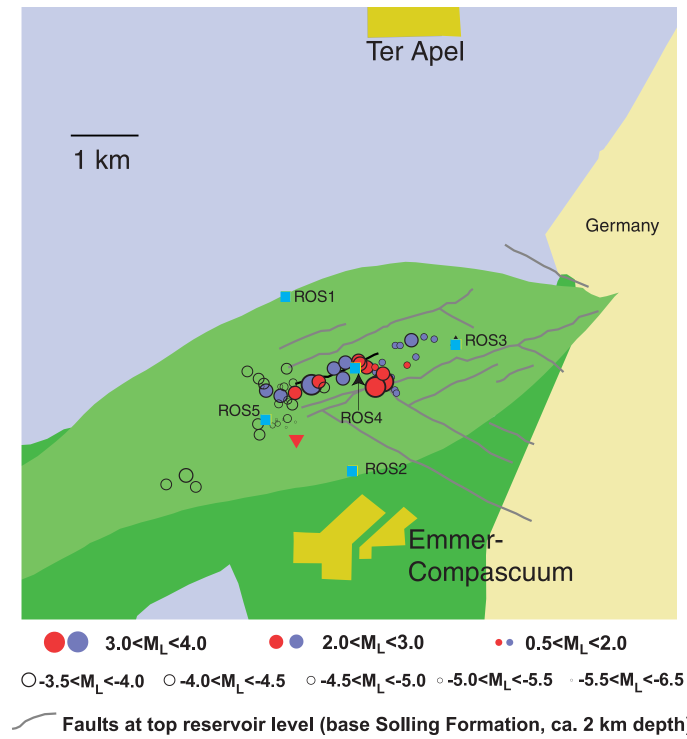

Fig. 4. Map of the Roswinkel area; see Fig. 3 for location. Gas fields in green. Blue squares indicate accelerometer stations (with station names); inverted red triangle indicates the micro-earthquake tool. Earthquakes are shown as filled circles; in red absolute locations using accelerometer stations, in purple relative locations. Open circles show results from the micro-earthquake survey.

reservoir itself. The Eleveld gas field was selected as a test case for the numerical modelling of this phenomenon, because all seismological information needed was available and the field is representative for several gas reservoirs in the area. Later, similar studies were made for the earthquakes near Roswinkel, where other mechanisms such as thrust or reverse faulting due to large horizontal pressures, are possible (Roest et al., 1998; Van Eijs, 1999).

In most cases, differential compaction is a gradual process that follows the drop in reservoir pressure. When fault movement in compacting rock is obstructed, possibly due to the presence of conglomerate, large shear stresses can build up. When these stresses exceed a threshold value, the obstruction will be broken and an earthquake results. Following McGarr & Simpson (1997) this mechanism is referred to as induced. It is clear that in terms of stress, the pressure drop within the reservoir is the driving force behind the induced earthquakes, in terms of strain it is the compaction. One might expect that for a producing reservoir which is at least in part fault closed and which, because of its large diameter/depth ratio D/d, shows no significant extension in its overburden, earthquakes should only occur at the faulted edges of the field. This, however, is not always the case. In the large Gronin-

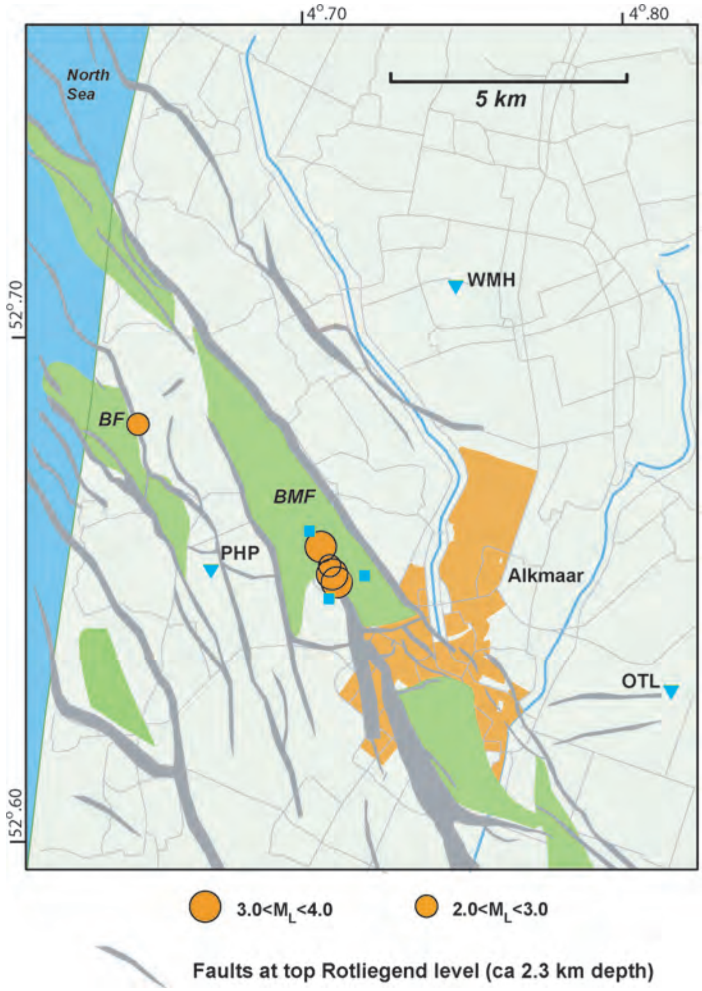

Fig. 5. Map of the region around Alkmaar, Noord-Holland. Borehole seismograph stations are shown as inverted blue triangles, accelerometer stations as blue squares. Gas fields are in green and faults at top Rotliegend in grey. Fault width is exaggerated, showing fault dip over the depth of the Rotliegend. Gas fields: BF: Bergen; BMF: Bergermeer.

gen field for example, where $D/d > 10$, and where the compaction is relatively small, numerous earthquakes are induced within the field. They are concentrated along large faults at reservoir level (Rotliegend, just below base Zechstein, Fig. 3), and clearly result from differential compaction between some of the many interconnected producing compartments that constitute the field.

Overall, a wide variety in size and state of production of gas reservoirs is observed in the Netherlands. In the last 15 years only a minority of these reservoirs showed seismic activity, especially in the north-east. This regional differentiation of the induced seismicity is yet to be explained.

Induced seismic hazard and risk

Like for natural earthquakes, it is also important to know the magnitude of the most severe induced event that can be expected in a certain period as well as which damage might result from it. The dataset of events recorded by the borehole network forms the basis of the statistical calculations concerning the maximum expected magnitude and detection threshold. The number of recorded events, however, is limited, and estimates of the maximum expected magnitude improved as this number increased over the years.

Based on the seismicity recorded for the region where induced events occur, De Crook et al. (1998) show a 90% probability that earthquakes caused by gas extraction will not exceed magnitude $M_L = 3.8$. In the case of natural earthquakes, large fault zones usually enable large movements. In the case of induced events, however, it is not clear whether the existing faults are sufficiently large to enable the calculated maximum expected magnitude. According to the model of Brune (1968), the seismic moment of an earthquake, i.e. the physical measure of its strength, is the product of the relative displacement at its focus, the surface area of the ruptured fault, and the shear modulus of the local rocks. When volume changes occur a similar formulation applies (McGarr, 1976). The magnitude of the earthquake will be limited by geometrical boundary conditions close to the ground surface.

From calculations based on Brune's model, and on assumptions on fault-plane dimensions derived from 3D-seismic surveys, it can be deduced that the strength of an earthquake caused by gas production in the north of the Netherlands will not likely exceed a magnitude of about $M_L = 3.8$. This agrees with the results from the calculations based on recorded seismicity mentioned above. Figure 6 shows the frequency–magnitude relation for the north of the Netherlands. Two regimes can be distinguished. The standard low-magnitude regime has a *b*-value of 0.8. The higher-magnitude regime can be interpreted as either an exponential function leading asymptotically to a maximum magnitude, or as a second linear dependence. The higher *b*-value of the latter, indicating an increased number of smaller events with respect to the larger events, could be caused by a limited fault depth resulting in a more one-dimensional behaviour of the fault systems involved (Rundle, 1993; Rolando et al., 1997).

Since seismological instrumentation was lacking when the first induced earthquakes were felt by people, macroseismic information was gathered by means of public inquiries. Based on these data, intensities were calculated on a regular grid in the epicentral area. Intensity maps, showing isoseisms, were used to evaluate the area of possible damage, and to infer the depth of the earthquake. The estimated maximum intensity of earthquakes in the north of the Netherlands will reach about intensity VI-VII. The effects of intensity VI first became apparent in the area of Roswinkel, where an earthquake of magnitude $M_L = 3.4$

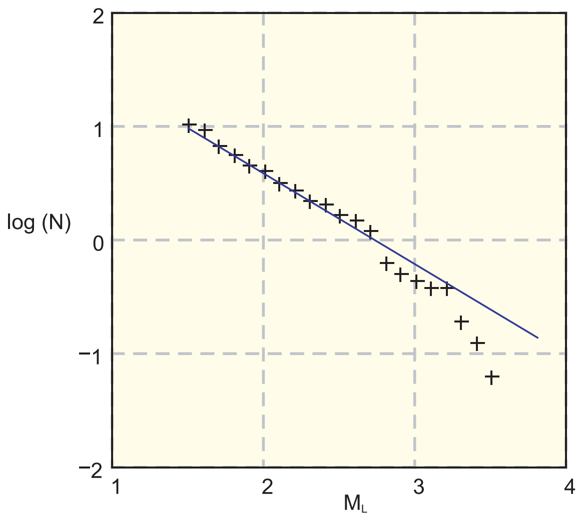

Fig. 6. Logarithm of the cumulative annual frequency of the total dataset of induced events, part of which is listed in Table 4, as a function of magnitude (M_L). Measurements (+) and best fitting regression, $b = 0.8$, for the linear part of the curve are shown.

at a depth of only 2 km occurred on February 19, 1997 (Dost & Haak, 1997). More than 200 cases of damage were reported in an area with a radius of 5 to 6 km from the epicentre. The damage ranged from broken crockery and glassware to small cracks in walls. No structural damage was reported. Above intensity VI, by definition, more serious structural damage to buildings is to be expected (Gruenthal, 1998). Intensity VI corresponds to induced quakes of roughly magnitude $M_L = 3.5$ at depths of 2 to 3 km, as well as to tectonic quakes at 15 km depth of roughly magnitude $M_L = 4.5$.

The estimation of intensities and horizontal peak accelerations from instrumental measurements proved to be useful for the objective determination of the possible damage after stronger events. During 1996, five accelerometers were placed in the north of the Netherlands at locations that experienced several events (Fig. 3). In these locations, chances were high to catch an event close to its epicentre, and to estimate its epicentral acceleration with instrumental means. Until December 2002, 21 events were recorded near Roswinkel by one or more of the accelerometer stations, showing a maximum acceleration of 3 m/s^2 for the February 19, 1997, $M_L = 3.4$ event at a hypocentral distance of 2 to 3 km from the epicentre (Figs 4, 7). The dominant frequency of the S-wave is around 10 Hz, implying a short duration of the large peak acceleration. After this largest Roswinkel event, two additional instruments were installed, followed in 1999 and 2000 by two more. The existence of high seismic intensities proved to be right. Extension of the data set will allow an improved analysis of the relation between maximum intensity and magnitude of future earthquakes induced by gas production.

The effects in terms of damage depend to a great degree on the construction methods used for houses and buildings and on their histories of use. The use of brick construction materials combined with changing ground-

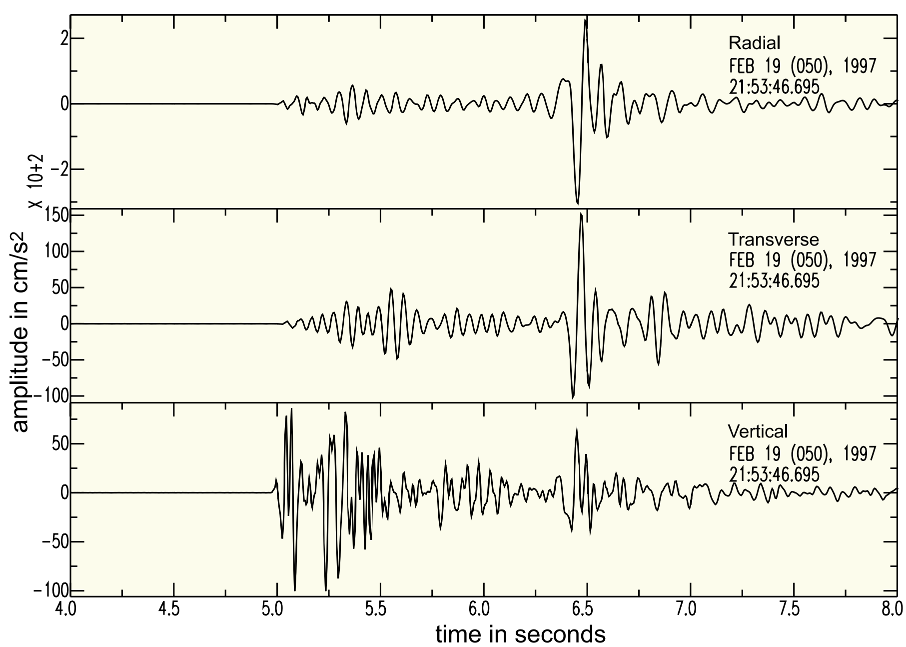

Fig. 7. Acceleration record at 2.5 km hypocentral distance from the February 19, 1997 Roswinkel event ($M_L = 3.4$). The maximum horizontal acceleration of the radial component amounts to 3 m/s^2.

water levels led for example to pre-existing cracks in buildings that may have been widened during the earthquakes.

In some instances people report that they felt an earthquake outside the region where one would expect this on the basis of instrumental observations and other felt reports. This may be due to differences in shallow geology (0-30 m). For example, under certain conditions, slightly consolidated sediments can amplify the vibrations at the surface by resonance effects. From one-dimensional and non-linear modelling exercises, using vertically up-going S-waves (e.g. Schnabel et al., 1972), and also using field measurements of seismic velocities, densities and damping values, it became clear that this presents a real effect. The model calculations agree well with measured earthquake data at an experimental site (ZLV) near Zuidlaarderveen, NE of Assen, which is equipped with a borehole string of 3-component geophones, 25 m apart, to a depth of 200 m. This system could follow the amplitudes of the up-going wave fronts of nearby earthquakes (De Crook & Wassing, 2001).

Conclusions

In a low-seismicity area like the Netherlands, where moderate earthquakes occur at recurrence intervals of 50 to 100 years, long-term continuous monitoring is required to study details of fault behaviour. Due to the increased station quality and density installed over the past 20 to 30 years, the earthquake sources can now be studied in detail, not only in the south of the country where the seismicity is natural, but also for the events induced in the north. In both regions the seismicity is related to faults. In the induced events, normal faults are reactivated in reverse sense as a result of the removal of gas from a reservoir. In both regions the damage reported is caused by the largest events. The difference in average earthquake depths between the regions, however, makes it difficult to define a threshold magnitude at which damage starts to occur. For the shallow, induced events this magnitude may be $M_L = 3.2$ or more, while for the natural events it is around $M_L = 4.5$. Quantification of the peak horizontal accelerations in the epicentral regions, and the study of site effects are the next steps in the evaluation of seismic hazard and risk.

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Petroleum geology

J. de Jager & M.C. Geluk

ABSTRACT

A great variety of proven hydrocarbon plays and trap styles are present in the sedimentary succession of the Netherlands, which starts in the middle Paleozoic and has been deformed by several periods of structuration. The thick Permian Zechstein salt in much of the subsurface provides an effective seal between a prolific Paleozoic gas system and a mixed, oil and gas-prone Mesozoic hydrocarbon system. The giant Groningen field represents two thirds of the recoverable Dutch gas reserves. It contained an initially recoverable gas volume of ca. 2700 × 10^9 m^3 in thick, mainly fluvial sandstones of the Permian Upper Rotliegend Group. The Rotliegend play is formed by a near-ideal superposition of i) the thick Upper Carboniferous, Westphalian succession with abundant coal measures as source rocks for gas, ii) good Rotliegend reservoir sandstones, and iii) the excellent seal of the Zechstein salt. Triassic plays are second in importance with respect to proven gas volumes. Additional gas reserves are in Permian Zechstein carbonates, Jurassic and Cretaceous sandstones, Upper Cretaceous chalk as well as in shallow unconsolidated sands of Tertiary and Quaternary age. Producible oil occurs within the Late Jurassic and Early Cretaceous rift basins in a variety of sandstone reservoirs and trap styles. Only minor amounts of oil have so far been found in the Upper Cretaceous chalk. The exploration of the Dutch hydrocarbon plays is in a mature stage. Yet, new discoveries continue to be made, and assessments of remaining potential show that significant volumes, in particular of gas, are yet to be found.

Keywords: Netherlands, hydrocarbon plays, exploration, production, reserves, remaining potential

Introduction

The Netherlands is a gas country. Since the discovery of the giant Groningen field in 1959, the Netherlands has developed into the main gas 'hub' of western Europe. While before the 1960s gas was considered to represent a risk in exploration – one would rather find oil – it is now the preferred fossil energy source. Moreover, the Netherlands has its share of oil as well. The Schoonebeek oil field, now closed-in and only producing from its extension into Germany, is the largest onshore oil field of western Europe. Other oil fields occur in the western Netherlands and in the offshore (Fig. 1).

Most gas is contained in the Upper Permian, Upper Rotliegend sandstones. Other hydrocarbon plays, ranging from Carboniferous to Quaternary, are present as well. The discovery of the Groningen field close to a large potential market triggered an intense exploration and production activity both on- and offshore. As a result, a wealth of new data has been acquired over the last few decades. Currently, a total of ca. 56 000 km^2, ca. 56% of the total Dutch on- and offshore surface, is covered with high-quality 3D seismic data, allowing a good appreciation of the great variety of geological settings and structural styles (Fig. 2). Indeed, the Netherlands must rank amongst the countries that have the best seismically imaged subsurface geology in the world (Dessens, 1996). In addition, by January 1st 2006, the oil industry had drilled a total of 1105 exploration wells (Fig. 3), 403 appraisal wells and 1566 production wells (Ministry of Economic Affairs, 2006).

Exploration and production history

A brief summary of the history of oil and gas exploration in the Netherlands follows below. Reference is made to the more extensive overviews published by Knaap & Coenen (1987), Breunese & Rispens (1996) and Glennie (2001), and also to the yearly review by the Ministry of Economic Affairs, under the title 'Oil and Gas in the Netherlands, Exploration and Production' (available from http://www.nitg.tno.nl/oil&gas).

The first natural oil in the Netherlands was found in 1923, near Corle, close to the German border, when 1.5 barrels of oil (240 l) were recovered from Zechstein anhydrites and Carboniferous sandstones (Fig. 4). A blowout in 1938 at Bad Bentheim in Germany, ca. 50 km NE of Corle, indicated the presence of gas at Zechstein level (Knaap & Coenen, 1987). The second indication for oil was found in 1938, during an exhibition in The Hague, where the Bataafsche Petroleum Maatschappij, the predecessor of Shell Internationale Petroleum Maatschappij, drilled a demonstration well. At a depth of 460 m, the drillers unexpectedly encountered oil stains. The first commercial oil was found five years later, in 1943, when the Schoonebeek field was discovered in Lower Cretaceous sandstones at a depth of ca. 800 m in the south-east of the province of Drenthe. With an initial in-place volume of ca. 1 billion (10^9) barrels (160 × 10^6 m^3), this is the largest onshore oil field of western Europe. The field became uneconomic when 25% of the viscous oil (25° API) had been produced, and is closed-in since 1996. NAM, the Neder-

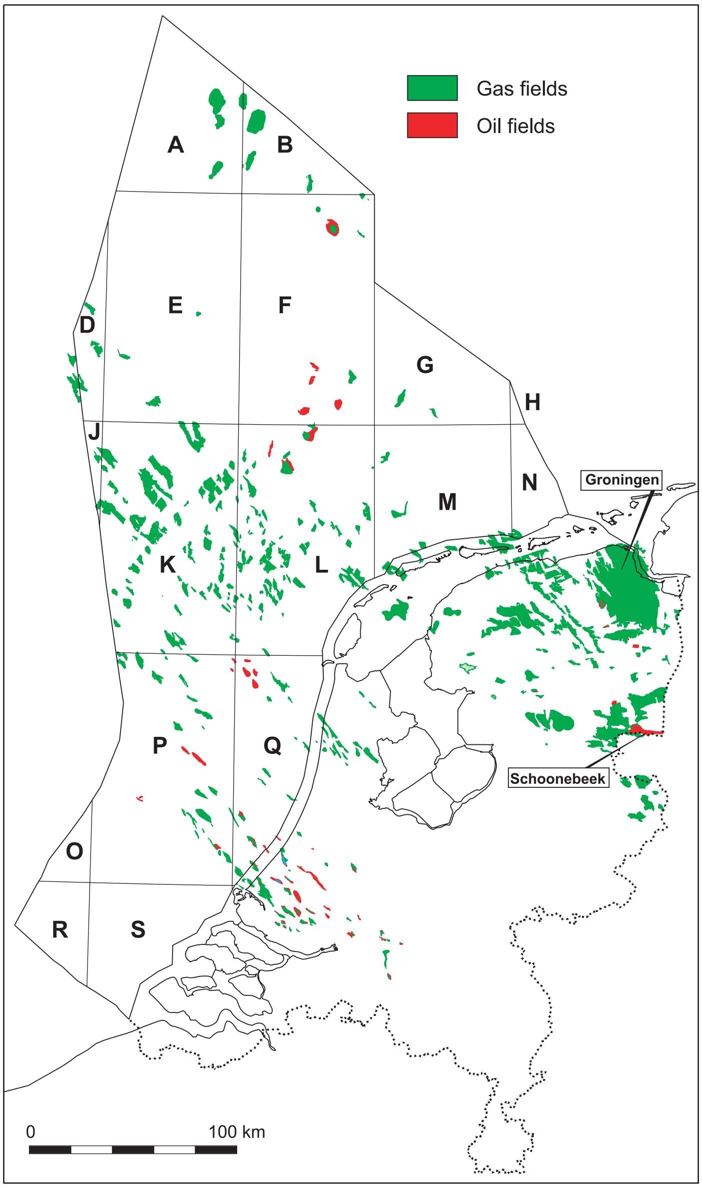


Fig. 1. Gas and oil fields in the Netherlands. Total reserves are dominated by the giant Groningen gas field in the north-east onshore.

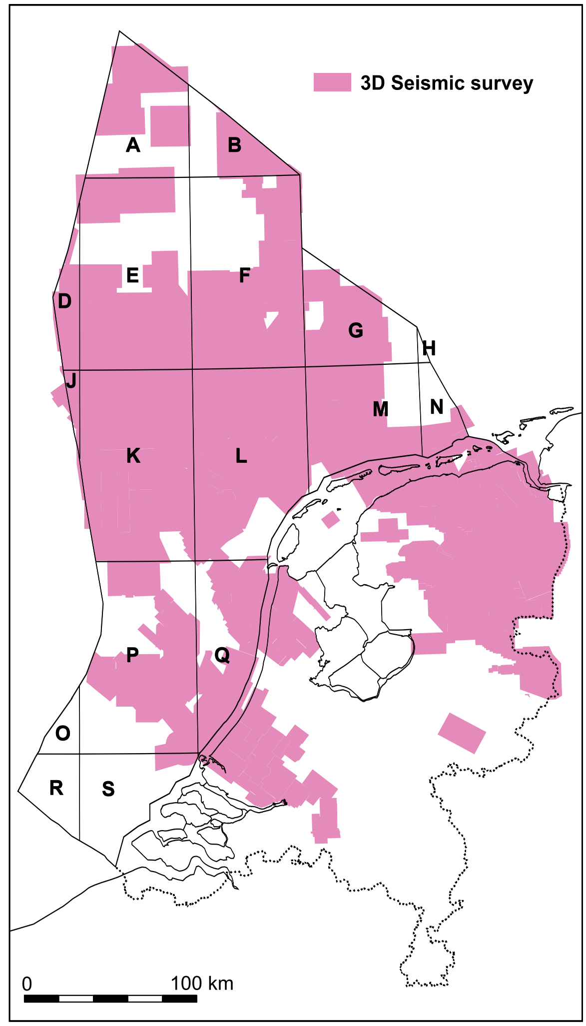


Fig. 2. 3D seismic coverage. In total ca. 56 000 km² of 3D seismic has been acquired, representing ca. 56% of the Dutch on- and offshore areas.

landse Aardolie Maatschappij, currently studies redevelopment of this field.

After World War II, exploration resumed with further oil finds in the West Netherlands Basin (e.g. the Rijswijk, IJsselmonde, Wassenaar, Ridderkerk and Rotterdam fields) and gas discoveries in the east of the country (e.g. the Coevorden, De Wijk, Wanneperveen and Tubbergen fields). Until the end of the 1950s the Zechstein carbonates and Lower Cretaceous sandstones were considered the most prospective reservoirs. The giant Groningen gas field, which would change the exploration outlook for good, was discovered in 1959 with the Slochteren-1 well. It has been reported many times that the sheer size of this field was initially not recognised. The well's target was a relatively small structural closure at the level of the basal Zechstein carbonates, in which gas had already been encountered in south-east Drenthe. However, the sandstones at approximately 3 km depth in the underlying Permian, now the Slochteren Formation, turned out to contain gas as well. Only several wells later it was realised that the small structure that was targeted forms part of a much larger structure with a length of more than 40 and a width of almost 30 km. With initial recoverable reserves of ca. 2700×10^9 m^3 of gas, the Groningen gas field is by far the largest of Europe. It represents two thirds of the total initially recoverable proven gas volumes in the Netherlands.

In the early 1960s the first careful steps into the offshore were made with the near-coastal well Kijkduin Zee-1. Following legislation for the offshore in the mid 60s, the first gas was discovered in 1968 in block P6, and the first oil in 1970 in block F18. Offshore gas production started in 1975 with the large L10 Rotliegend gas field, and the first

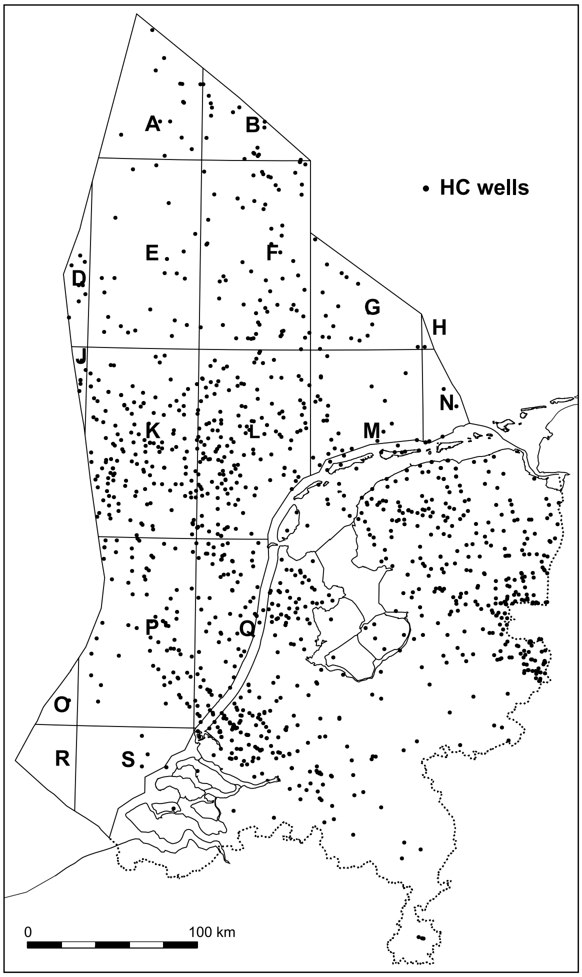


Fig. 3. Hydrocarbon exploration and appraisal wells. A total of 1105 exploration wells and 403 appraisal wells had been drilled by the end of 2005.

offshore oil was produced in 1982 from the Cretaceous in block Q1. The application of 3D seismic since the 1980s has resulted in a much increased exploration efficiency. Currently ca. 50% of the Dutch on- and offshore areas has been awarded as exploration or production licences (Fig. 5).

Legislation

On January 1st, 2003 a new Mining Law (Mijnbouwwet) has become effective governing exploration and exploitation of oil and gas for both the onshore and offshore. This law recognises different licences for exploration, production, underground storage, mining and pipelines. With the application for permission to drill a new exploration well, a production plan must be submitted. Further details are contained in separate decrees and regulations. Data acquired during exploration and production remain confidential for a period of 5 years, after which they are released to the public. All data ac-

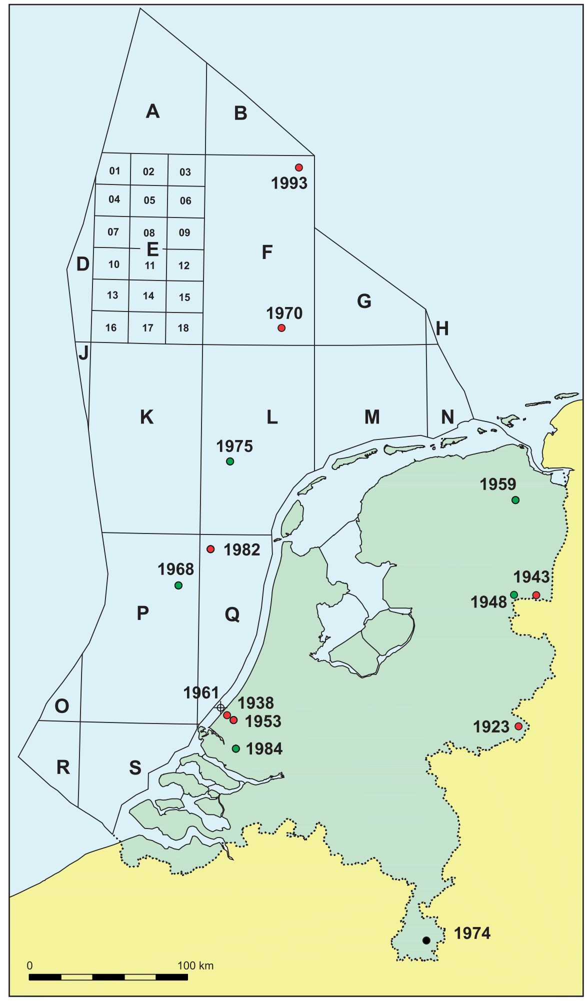


Fig. 4. Milestones in the Dutch exploration and production history (after PGK, 1993). 1923: Corle oil show; 1938: The Hague oil show; 1943: Schoonebeek oil discovery (on stream in 1945); 1948: Coevorden gas discovery (first production in 1951); 1953: Rijswijk oil discovery (first production in 1954); 1959: Slochteren gas discovery (Groningen field on stream in 1963); 1961: First offshore well (Kijkduin Zee-1); 1968: First offshore gas discovery (P6); 1970: First offshore oil discovery (F18); 1974: Last coal mine in Zuid-Limburg closed; 1975: First offshore gas production (L10); 1982: First offshore oil production (Q1); 1984: Botlek gas discovery in Rijswijk concession; 1993: F3 oil field on stream.

quired prior to 2003 will be released 10 years after acquisition.

Production

The cumulative production until January 1st, 2006 in the Netherlands amounted to 2839×10^9 m^3 of gas and 832 million barrels (128×10^6 m^3) of oil, with remaining proven reserves of 1510×10^9 m^3 of gas and 234 million barrels (36×10^6 m^3) of oil. In 2005 there were 188 producing gas fields, good for a yearly production of some

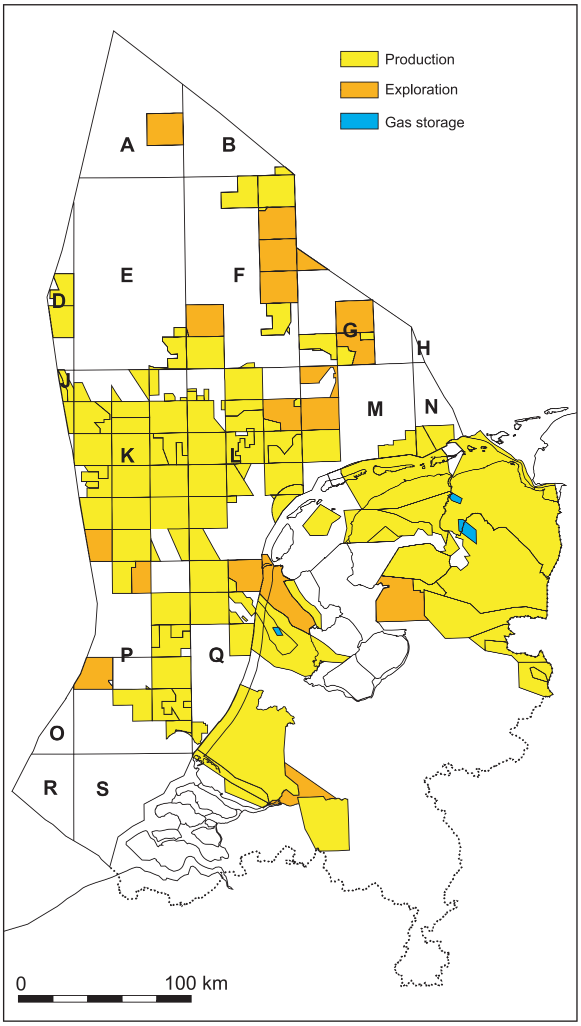


Fig. 5. Licence situation for hydrocarbon exploration and production in the year 2005.

73×10^9 m³, while the Dutch gas consumption is some 50×10^9 m³ per year. The Netherlands is consequently a net gas exporter, with annual export of some 30×10^9 m³ and import of 10×10^9 m³. The annual oil production of almost 9.9 million barrels (1.53×10^6 m³) from 13 fields is relatively modest. In the year 2005, the total hydrocarbon production provided the Dutch State with an income of €7.5 billion (Ministry of Economic Affairs, 2006). In order to save gas reserves for future generations, the State has set a ceiling for yearly domestic gas production. In the same light there are tax incentives for the development of gas reserves in small fields ('kleine velden beleid' = small fields policy). The small fields are allowed to produce at their optimal production rates, while the large Groningen field is used to provide extra production capacity as swing producer during winter. With the ongoing production and corresponding pressure decline in this field, it will become increasingly difficult to fulfil this role during extreme cold-weather spells. Three underground gas storage facilities (Grijpskerk, Langelo and Alkmaar) have therefore been installed to provide additional swing production capacity (Bos, this volume).

In an economic sense, a distinction is made in the Netherlands between two gas types: gas with a lower heating value ('low cal' or Groningen quality), generally with a nitrogen content of more than 10 to 12%, and gas with a high heating value ('high cal'), containing less non-hydrocarbon gas. There are separate pipeline systems for both gas types. The low-cal gas is being distributed to Dutch households, while high-cal gas is mainly for export and industrial use. An overview of the composition of the gases in the Netherlands is presented in the Northwest European Gas Atlas (Lokhorst, 1998).

The petroleum industry and society

In the densely populated and low-lying country of the Netherlands, there is a concern that exploration and production activities may negatively influence the environment, or the well-being or comfort of the people. The main issues are subsidence, earth tremors, and operations in sensitive areas.

SUBSIDENCE

Subsidence of the surface affects the surface water runoff and the groundwater table, and is widely felt to be a serious issue. Mining activities, including the production of oil or gas, can cause such subsidence. Other processes that can result in subsidence include tectonics, compaction of near-surface layers and (man-induced) lowering of the groundwater table. Subsidence caused by oil or gas production is monitored at regular intervals and the environmental implications are evaluated. Prior to the start of production, sometimes even in the exploration phase, predictions of future subsidence are issued for all onshore oil and gas fields.

At the surface above the Groningen field, which has been in production from 1963 onwards, to date a maximum subsidence of some 26 cm has occurred in the centre of a 'bowl' with a diameter of more than 40 km. The maximum subsidence after gas production is expected to be less than 45 cm (NAM, 2000).

At the surface above the Ameland field, i.e. on the island of Ameland, in the Waddenzee area and in the adjacent coastal zone of the North Sea, the maximum subsidence after 15 years of gas production is ca. 25 cm. However, most of the subsidence in the Waddenzee as well as in the marsh areas on the island has been compensated by natural sedimentation. As a result the impact on the environment has been negligible and the ecological value of the area has not been affected (Eysink et al., 2000). All new gas fields to be developed in the area would be smaller and cause less subsidence, rarely more than 10 cm in the centre of the subsidence bowl.

Occasionally the predicted subsidence that would result from gas production may lead to a decision not to produce a certain accumulation, as even limited subsidence may have undesirable implications locally. The issue of subsidence also played a key role in the discussions about the development of new gas fields in the Waddenzee (see below).

It must be noted that subsidence resulting from lowering of the groundwater table is often greater and more wide-spread than subsidence caused by gas production, particularly in peat areas. For example, in the province of Friesland, away from producing gas fields, it is predicted that an area of 80 km^2 will have subsided more than 40 cm by the year 2050 (Province of Friesland, 1997).

Earth tremors

Variations in subsurface stress fields as a result of production-induced pressure reduction can lead to earth tremors. This has occurred in the north-east Netherlands, e.g. above the Roswinkel field, and in the province of Noord-Holland, e.g. above the Bergermeer field. Most of these tremors have a magnitude of not more than 2 on the Richter scale. Such tremors can in extreme cases cause rattling of doors and windows and sometimes minor damage. The maximum-recorded tremor had a magnitude of 3.5 (Dost & Haak, this volume). The maximum magnitude that could occur through gas production will probably not exceed 3.8 (De Crook et al., 1998), and could cause structural damage to houses near the epicentre. In general the responsible oil company compensates any damage that has been ascribed to production-induced earth tremors by independent experts, or by a committee installed by the Ministry of Economic Affairs.

Operations in sensitive areas

The Mining Law stipulates the required procedures and permits for exploration and production operations. Other laws also play a role in the permitting and execution of these operations: the Wet Ruimtelijke Ordening (spatial planning law), Wet Milieubeheer (environmental law) and the Natuurbeschermingswet (nature protection law). Permitting requirements vary depending on the nature of the area. Onshore, the areas with the highest level of protection include ecologically designated areas, 'silence-preservation areas', soil-conservation areas, bird- and habitat-directive areas and nature-protection areas. Offshore, restrictions apply in ecologically designated areas, shipping lanes and military exercise areas. For all new drilling or production locations onshore a building permit must be obtained from the local authorities. For activities in sensitive areas a location permit must be obtained from the Ministry of Economic Affairs, coupled to an environmental impact assessment (in Dutch: MER, Milieu Effect Rapportage). Other permits that could be required include nature-protection and environmental permits. During all permitting procedures, civilians, non-government organisations or other organisations may file objections. For activities in or near sensitive areas it may take several years before all permits are secured. The new Mining Law, effective January 1st, 2003, considerably affects the requirements for exploration and production permits. The main new elements are that for all exploration wells an environmental permit from the Ministry of Economic Affairs is needed and that production may only commence (or proceed in the case of fields already developed) after consent by the Ministry to the submitted production plan (comparable with a condensed field development plan). Together with the application for a production permit the operator must submit a field development plan.

The exploration and production of gas from below the Waddenzee has met strong objections and considerable media attention. Currently, there are two gas fields in production in this area: Ameland and Zuidwal, both since the mid 1980s. Subsequent exploration drilling for gas prospects below the Waddenzee from onshore surface locations resulted in the discovery of some 40×10^9 m^3 of gas in new fields. In 1999, however, the government decided not to allow production of this gas nor further exploration activities, pending the outcome of studies of potential environmental damage through subsidence. Following these studies, Parliament endorsed gas production in November 2004.

Petroleum systems

In the Dutch subsurface the gas plays are volumetrically and economically by far the most important. With respect to the ages of source rocks, reservoirs and seals these plays belong predominantly to a Paleozoic hydrocarbon system. Where the thick Permian Zechstein salt is present, it provides an effective seal between this system and the oil plays which belong almost entirely to Mesozoic hydrocarbon systems (Fig. 6).

Gas: source rocks and generation

The principal source rocks for gas are the Upper Carboniferous, Westphalian coals and carbonaceous shales, which are present in much of the subsurface. Almost all the gas found has been generated from these source rocks (Lokhorst, 1998; Gerling et al., 1999). The cumulative thickness of the coals is several tens of metres. They occur mostly in the Maurits Formation (Westphalian B), and are less common in other Westphalian units. Because of Early Permian uplift and erosion the Westphalian source-rock thickness is locally much reduced. Where the total Westphalian of ca. 5.5 km thickness is preserved, the ma-

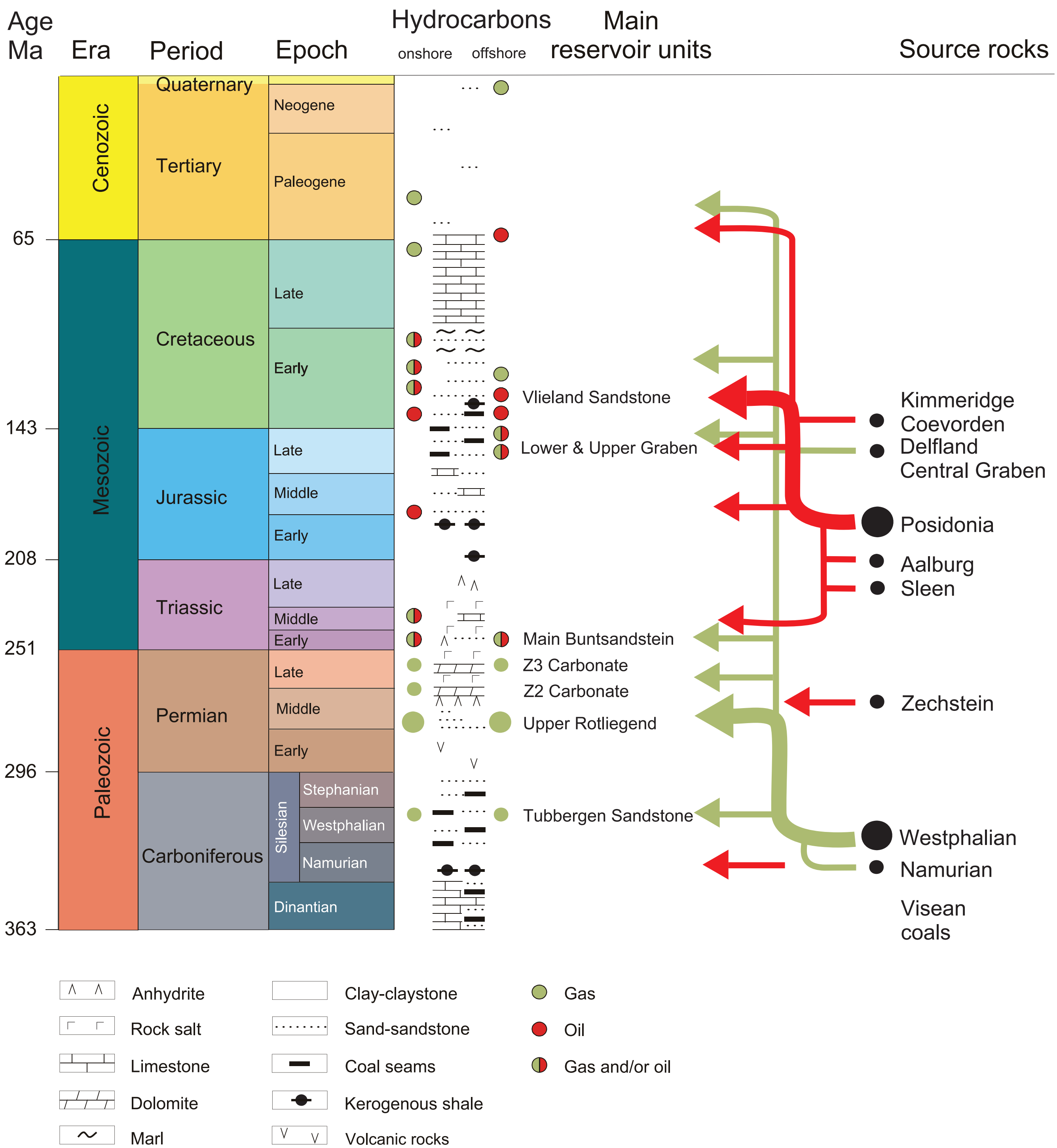


Fig. 6. Hydrocarbon systems in the Dutch subsurface. Arrows show from which source rocks the main reservoirs have been charged with gas and/or oil. The Upper Permian Zechstein salt, present in much of the subsurface, provides a regional seal between a Paleozoic gas and a Mesozoic oil and gas system. Not shown is that with time probably 98% of the generated hydrocarbons escaped into the biosphere.

turities vary significantly from top to bottom. Secondary source rocks for gas occur in basal Namurian organic-rich shales (Lokhorst, 1998; NITG, 1998; Gerling et al., 1999). In most places these source rocks became over-cooked during deep pre-Kimmerian burial. Nevertheless, the Namurian is thought to have contributed significantly to the nitrogen charge, which is mainly expelled at much higher temperatures than hydrocarbon gas.

In general, hydrocarbon generation from the Westphalian coals was widespread until the Middle Jurassic. After the Middle Jurassic, a distinction must be made between the Kimmerian rift basins and the platforms and highs. During the Late Jurassic to Early Cretaceous rifting, hydrocarbon generation accelerated within the rift basins as a result of increased subsidence. This generation halted during the Late Cretaceous due to inversion-related uplift and declining heat flow. At the margins of the basins, where inversion had been limited and was followed by strong Tertiary subsidence, for example on the south-west margin of the West Netherlands Basin, charge from the Westphalian resumed during the Tertiary and continues until the present day (De Jager et al., 1996). The platforms and highs, on the other hand, were uplifted during the Late Jurassic, interrupting hydrocarbon generation. Where subsequent burial caused temperatures at the Westphalian source-rock levels to exceed the maximum temperatures reached earlier, gas generation resumed.

Secondary source rocks for gas occur in Upper Jurassic and Lower Cretaceous coals of the Delfland Subgroup in the West Netherlands and Broad Fourteens basins, and of the Central Graben Subgroup of the Dutch Central Graben and Terschelling Basin. Furthermore, Upper Dinantian, Visean coals, like those encountered in wells A16-1 and E6-1, may be relied on for some gas charge in the northern offshore, provided the timing of expulsion has not been too early.

Gas compositions

The gas quality in the various reservoirs shows distinct variations (Lokhorst, 1998). The most abundant non-hydrocarbon component is nitrogen. Differences in nitrogen content, from almost zero to more than several tens of percent, can be a result of differences in source rocks, but also of differences in their heat-flow and burial histories. Until the Late Jurassic, heat-flow rates were high and the Westphalian coals were expelling hydrocarbon gas while the much deeper Namurian shales expelled nitrogen (Fig. 7). This phase of charge was therefore in most places relatively rich in nitrogen. It ceased during times of uplift and significant erosion, e.g. during the Late Jurassic for the platform areas and during the Late Cretaceous and Early Tertiary for the rift basins. During subsequent burial, temperatures increased again, but under a lower heat flow than during the Late Jurassic (Fig. 8). Consequently, temperatures may today exceed maximum paleo-temperatures at the shallower Westphalian levels, while this is not (yet) the case at the deeper Namurian levels. The gas expelled during the Tertiary from the Westphalian is therefore not diluted by nitrogen from these deeper levels. Thus, in areas with present-day charge, gas fields generally contain less nitrogen than in areas with only 'old' charge.

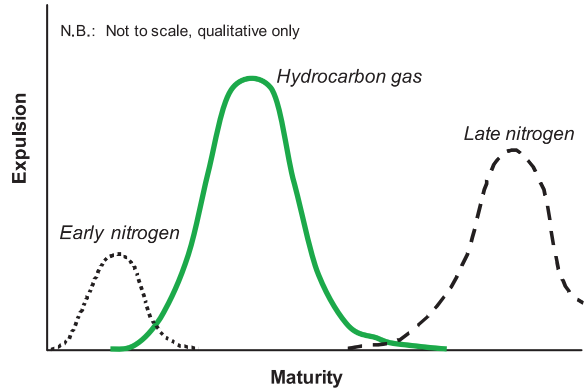


Fig. 7. Expulsion of hydrocarbon gas and nitrogen versus maturity. Apart from a minor early phase of generation, most nitrogen is expelled at much higher maturities than hydrocarbon gas.

Gas accumulations in the Carboniferous on the Cleaver Bank High contain more nitrogen (2–20%) than those in the east Netherlands (0–10%), and are in general 'wetter'. The gas with the anomalously high nitrogen content of 30% in block E12 was probably generated from Namurian source rocks. The CO_2 content in the Carboniferous reservoirs is locally as high as 10%. Methane carbon-isotope ratios are around –21‰ on the Cleaver Bank High and between –33 to –36‰ in the east Netherlands, probably due to lower source-rock maturities on the Cleaver Bank High, as observed in vitrinite reflectances. However, differences in source-rock type could also play a role (Gerling et al., 1999).

Gas in Rotliegend reservoirs is mostly wet, with wetness ratios ($C_1/(C_2 + C_3)$) varying between 10 and 25. The regional westward decrease of nitrogen, from more than 15% to less than 5%, displays considerable local variations. The Groningen field for example, with 15% nitrogen, is flanked to the west by gas fields with only a few percent of nitrogen, which have been charged during the Tertiary with low-nitrogen gas (see above). In the Ameland area, on the Friesland Platform and in the Broad Fourteens and Central Netherlands basins, more than 20% of nitrogen occurs locally. CO_2 is generally found in small amounts only, although in the Broad Fourteens Basin some accumulations contain 25% or more. This may be related to CO_2 generation from Paleozoic carbonates during deep, pre-inversion burial. This gas is also significantly dryer (wetness ratio between 100 and 200) than elsewhere, consistent with high maturities. Minute concentrations of mercury and helium have been reported from the Groningen field (Stheeman, 1963; Morrison, 1972; Ronteltap, 1973).

The composition of gas in Zechstein reservoirs varies considerably, mainly because of locally high nitrogen con-

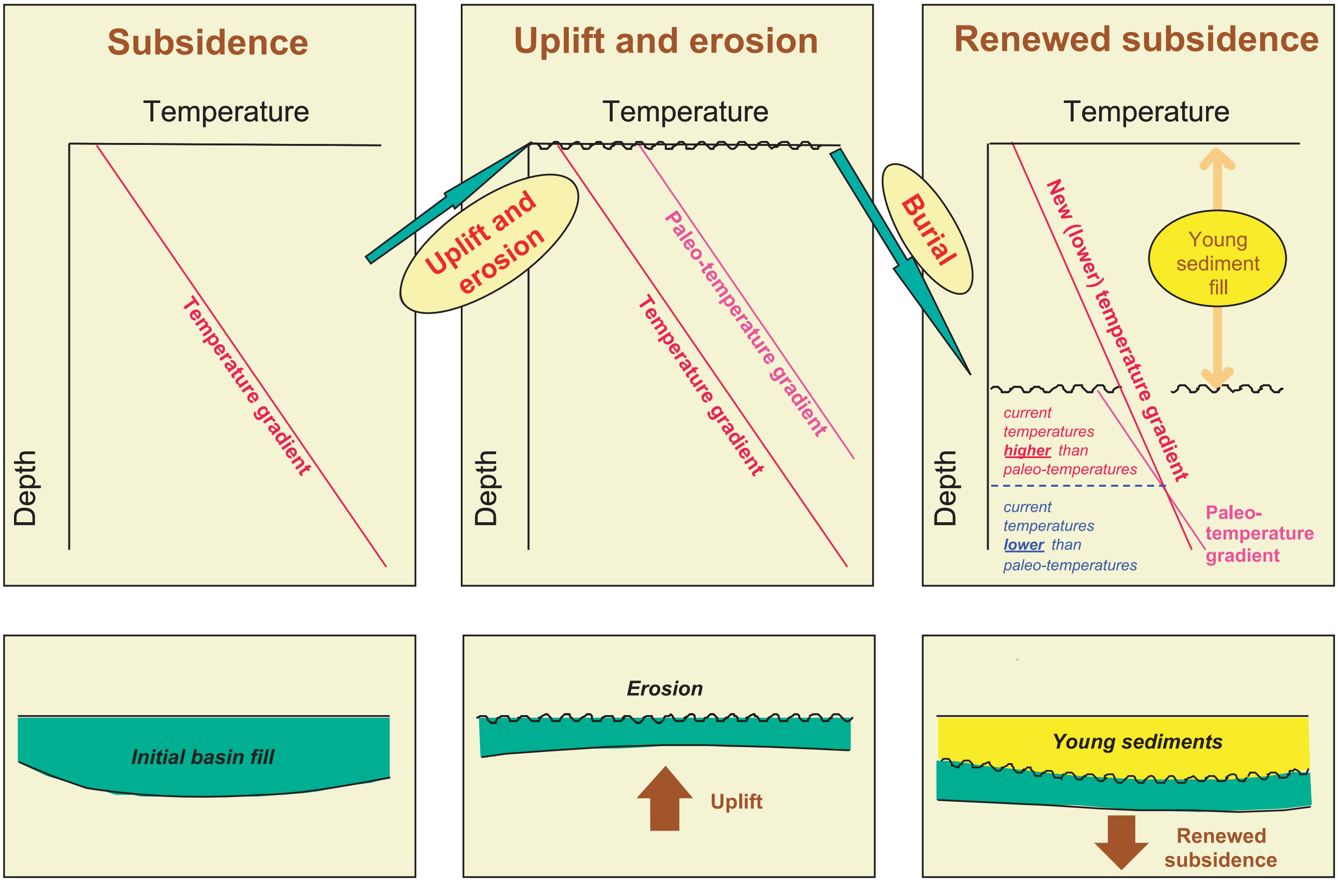


Fig. 8. Gas generation. In areas that have been uplifted and subsequently buried under a regime of declining heat flow, the shallower source rocks will be the first to reach higher temperatures than reached before uplift. Deeper source rocks will do so later. This has consequences for the wetness and nitrogen content of late charge.

tents, but also due to increased CO_2 percentages. In addition, many Zechstein gas accumulations in the eastern Netherlands are sour, and contain up to several tens of percent of H_2S (e.g. Vlagtwedde: 45%). This is a result of thermochemical sulphate reduction which strongly accelerates at temperatures above ca. 120°C (Orr, 1977). Consequently, significant quantities of sour gas occur only where the reservoir is buried sufficiently deep, and where anhydrite, required for sulphur supply, is present. However, the complex interplay of all controls on this reaction and other H_2S-generating processes, such as generation from carbonate source rocks and biochemical sulphate reduction, hampers accurate prediction of H_2S concentrations in undrilled prospects.

The gas in Triassic reservoirs generally contains less than 5% of nitrogen. Higher concentrations occur in the Broad Fourteens Basin (locally more than 20%), and also in the Dutch Central Graben (up to 10%). Anomalously high amounts of nitrogen are found locally in the east Netherlands (around 40% in Sleen and Roswinkel). Normally, CO_2 concentrations are less than 1%, while in the Dutch Central Graben they do not exceed 2%. The Werkendam field forms an exception with 70% of CO_2, probably related to local Jurassic volcanism. The gases in the Triassic are mostly wet, with wetness ratios of 25 to 50. In the Dutch Central Graben dry gas also occurs, with ratios of over 500. The De Wijk field also contains dry gas.

The gas in Jurassic and Cretaceous reservoirs contains 5 to 25% of nitrogen, and around 1% of CO_2. The gas is wet, with wetness ratios from 25 to 50. It is probably mainly sourced from the Carboniferous, with contributions in places from Jurassic coaly sequences, and possibly also from highly mature Posidonia Shale source rocks (De Jager et al., 1996).

The gas in Tertiary and Quaternary sands in the northern offshore differs markedly from the other gases. It is composed almost exclusively of methane. Methane carbon isotope data indicate a bacterial origin.

Oil: source rocks and generation

Unlike the areas to the north of the Dutch sector of the North Sea, the main source rock for oil in the Netherlands is not the Upper Jurassic Kimmeridge Clay. This elsewhere so prolific a source rock loses its kerogenous content just north of this sector. As in the Paris and Lower Saxony basins, the main Dutch source rock for oil occurs as rich, marine, Type-II source rocks in the Lower Jurassic, Toarcian, Posidonia Shale Formation. This source rock, which has only been preserved within the Late Jurassic rift basins, generated the oil that is trapped in the Upper Jurassic and Lower Cretaceous sandstone reservoirs in the West Netherlands and Broad Fourteens basins and the Dutch Central Graben (Fig. 9). It generally has a gross thickness of 15 to, locally, 35 m, with an average total organic carbon content (TOC) of ca. 10%, and a Hydrogen Index of up to 800. In strongly inverted sectors of the Kimmerian rift basins, the present-day temperatures of the Posidonia Shale may be lower than those reached before inversion. Where this is the case, the Posidonia Shale will not be generating oil at the present time.

The large Schoonebeek oil field was not charged from the Posidonia Shale, but from Lower Cretaceous lacustrine source rocks of the Coevorden Formation. These algal, Type-I source rocks are only known from the Lower Saxony Basin (Binot et al., 1991; NITG, 2000).

Additional source rocks for oil occur in the Lower Jurassic Aalburg and uppermost Triassic Sleen formations. They are of similar type as the Posidonia Shale, but less rich. Other source rocks for oil occur in the Permian Z2 Carbonate and Coppershale. While numerous oil shows in Zechstein carbonates are reported during drilling, both these source rocks have contributed only locally to oil accumulations (i.e. Stadskanaal, Gieterveen and E13-1). This is not only because of their limited thicknesses, but also because any oil from these source rocks trapped in Zechstein or Rotliegend reservoirs had a high chance to be flushed out by the subsequent abundant gas charge from the Westphalian. High condensate contents in Rotliegend gas can, however, often be correlated to Zechstein oil source-rocks.

Basal Namurian organic-rich shales with initially a high oil-generating potential have been encountered in deep wells (e.g. Winterswijk-1 in the eastern Netherlands) where, however, they are over-cooked. Based on regional paleogeographic settings, source rocks for oil may also occur in the Dinantian and Devonian (Cameron & Ziegler, 1997). Their great depths of burial would render them in most places over-mature for oil and gas generation.

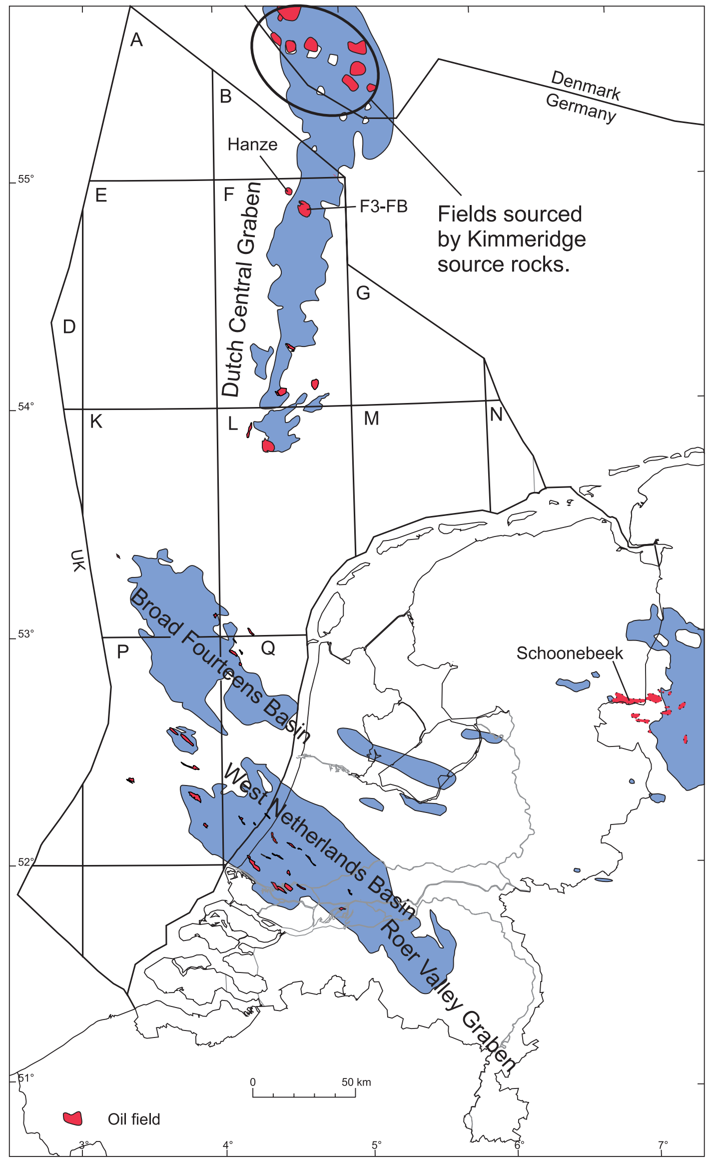


Fig. 9. The present-day distribution of the Posidonia Shale Formation (blue) is restricted to the Mesozoic rift basins, and clearly controlled the location of the oil fields. The large Schoonebeek field, however, was not sourced by the Posidonia Shale, but by lowermost Cretaceous source rocks (Coevorden Fm). The Danish oil fields in the north of the map were charged by Kimmeridge Clay source rocks.

Hydrocarbon plays

The proven gas and oil plays in the Netherlands range in age from Westphalian to Quaternary (Table 1). The Devonian and Lower Carboniferous are regarded as speculative because of the likely loss of reservoir quality and the post-maturity of their kitchens.

The Rotliegend play is by far the most important; the Groningen field accounts for two thirds of the Dutch initial gas reserves. Excluding Groningen, approximately 65% of all these reserves are in the Rotliegend. The total of proven initial Dutch gas reserves is 4349×10^9 m^3 (Ministry of Economic Affairs, 2006). For oil, there is a marked variation in the recovery factors (e.g. 12 to 40% in the West Netherlands Basin, 25% for Schoonebeek). For gas, the recovery factors generally vary between 70 and 90%, and may be above 95% (e.g. Groningen field). For gas in tight or very shallow reservoirs, or in reservoirs with low connectivity (e.g. in the Westphalian), recovery efficiencies

Table 1. Initial oil and gas reserves in the Netherlands at 1/1/2006.

Play	Initial gas reserves (10^9 m^3)	Initial oil reserves (10^6 m^3)
Westphalian	100	
Rotliegend–Groningen	2700	
Rotliegend–other	1000	
Zechstein	150	
Triassic	200	Minor
U. Jurassic & L. Cretaceous	100	Major
U. Cretaceous	5	Minor
Tertiary & Quaternary	20	
Total	4349	165

Numbers are based upon Ministry of Economic Affairs (2006), Geluk et al. (2002) and Geluk (1999). Note that as a result of rounding off, the total of reserves is slightly more than the sum of reserves per play.

are notably lower. Total initial oil reserves are reported to amount to just over 1 billion barrels (165 × 10^6 m^3) at US$ 55/barrel in early 2006. Production of several of the currently sub-economic oil accumulations becomes economically viable only at higher prices.

Westphalian play

The Westphalian is widely present in the Dutch subsurface, and is locally up to 5.5 km thick (Van Buggenum & Den Hartog Jager, this volume). Gas fields producing from Westphalian sandstones occur in the eastern Netherlands and adjacent parts of Germany, on the Cleaver Bank High in the north-western offshore and in the adjacent British areas (Fig. 10). Charge is derived from the Westphalian coals, with on the Cleaver Bank High some contribution from Namurian source rocks (Gerling et al., 1999). Traps are formed by dip-fault closures at the Base Permian Unconformity. Where Rotliegend sandstones overlie this unconformity, the Westphalian contains gas only where the height of the trap exceeds the thickness of the Rotliegend, e.g. in the Groningen field. Where no Rotliegend sandstones are present, like in the east Netherlands, the Zechstein salt forms an ideal top seal. On the Cleaver Bank High the claystones and evaporites of the Silverpit Formation are sealing. Locally, intra-Westphalian shales and faults have been found sealing as well (Fig. 11).

A key uncertainty for the Westphalian play is the presence of good reservoir sandstones within the traps at the level of the Base Permian Unconformity. Due to the Early Permian uplift and erosion, different rocks subcrop this unconformity. A good understanding of facies distribution within the Westphalian and mapping of intra-Westphalian horizons is therefore a prerequisite to pursue this play (Quirk, 1993; Quirk & Aitken, 1997). Fluvial sandstone reservoirs occur at distinct stratigraphic levels within the Westphalian. In the eastern Netherlands, such sandstones mainly occur in the Westphalian C and D (Tubbergen Fm). They are gas-bearing in several fields. The largest of these is Coevorden, at a depth of some 2800 m, with initially approximately 30 × 10^9 m^3 gas in place. Other significant fields are Dalen, Tubbergen and Hardenberg. The sands were derived from the rising Variscan mountain belt in the south-east, and the overall reservoir potential decreases towards the north-west. These typically sheet-like stacked sandstones have net-to-gross ratios as high as 50%. The uppermost Carboniferous (Westphalian D and Stephanian) comprises red beds, with low net-to-gross ratios of 10 to 20%. Porous conglomeratic sandstones occur, but their distribution is hard to predict. The reservoir architecture of the Westphalian sandstones and the faulted nature of the traps in the east Netherlands result in strong compartmentalisation of the gas fields.

On the Cleaver Bank High, gas-bearing sandstones occur mainly in the Westphalian A and B (Botney Mbr, Caister and Murdoch sandstones) and, similar to the Dutch onshore, Westphalian C and D (Hospital Ground Fm). The main fields are located in the northern K and southern D and E quadrants, at depths around 3500 to 4000 m, north of the pinch-out of the Rotliegend sandstones. The largest Westphalian field there is D15-FA, with an initial gas-in-place volume of ca. 5 × 10^9 m^3. The trend of Westphalian gas fields continues into the UK sector of the North Sea, with fields like Orca, Caister, Murdoch, Ketch and Schooner and various other discoveries (Mijnssen, 1997).

The Westphalian sandstones of the Cleaver Bank High were deposited by fluvial systems from northern source areas, and their reservoir quality deteriorates towards the south. This sandy facies may extend further east to the Schill Grund High, at the southern margin of the Ringkøbing-Fyn High, although there are insufficient well data to substantiate this. The Westphalian A and B deposits display generally low net-to-gross ratios.

Where the Westphalian sandstones are gas-bearing, they often show fair to good porosities (average 9%, maximum 20%) and permeabilities (average 1–2 mD, locally > 100 mD), and good initial production rates of about 1 × 10^6 m^3/day or more can generally be achieved (Van Buggenum & Den Hartog Jager, this volume). However, their poor reservoir connectivity often results in limited connected volumes per well and in rapidly declining flow rates, thus seriously hampering the economic viability of offshore Westphalian gas accumulations.

Rotliegend play

The Slochteren Formation of the Upper Rotliegend Group is, even excluding the giant Groningen field, volumetrically by far the most important gas reservoir in the Netherlands. The Rotliegend play is formed by an ideal superposition of i) the prolific Westphalian source rocks for gas, ii) the thick Slochteren sandstone reservoirs, and iii) the perfect seal of the Zechstein salt (Glennie & Provan, 1990). The Slochteren reservoir sandstones are well developed in the classical Rotliegend play which runs E-W from the Groningen field to the offshore K and L blocks. To the north these sandstones shale out into the Silverpit Formation. The southern boundary of the play is determined by the depositional limits of the Slochteren reservoir or of the Zechstein seal (Fig. 12). Most traps occur in simple horst blocks (Fig. 13).

The sands in the Upper Rotliegend were supplied mainly from the Variscan Mountains (Verdier, 1996; Geluk, this volume). They show marked differences in reservoir characteristics. Clean eolian and well-sorted sandstones form the best reservoirs, but there are also packages of up to tens of metres of mainly fluvial sandstones with, locally, conglomeratic intervals. The upper parts of the Slochteren sandstones shale out towards the north into the Ten Boer Member. Depending on the amount of sand present, this member acts either as a waste zone or as seal for the underlying Slochteren sandstones. Even where this member is not a top seal, it does provide a lateral seal when it is juxtaposed against Slochteren reservoirs (Fig. 14). Its sealing potential increases with increasing shaliness, which suggests that a sealing shale gouge formed along the fault planes. In some gas accumulations, the gas-water contact extends even below the mapped spill point given by the Ten Boer lateral seal, indicating that faults in clean sandstones can also be sealing. It is therefore likely that factors other than shale gouge, such as cataclasis and/or diagenesis, also play a role in the establishment of sealing faults (Leveille et al., 1997).

Below the Silverpit shales, basal Slochteren sandstones are locally present. Thickness variations of these sandstones are considered to be the result of paleo-topography, which is related to faulting and the subcrop below the Base Permian Unconformity. The shaly Westphalian A and B were less resistant to erosion than the sandy Westphalian C and D. The former formed lows, where sand accumulated; the latter formed ridges, where little or no sand was deposited (Geluk & Mijnlieff, 2001).

Most Rotliegend reserves occur in the north-east Netherlands and in the offshore K and L quadrants (Central Offshore Saddle, the southern sector of the Cleaver Bank High and the northern sector of the Broad Fourteens Basin). This trend of Rotliegend fields continues westwards in the British Indefatigable and Sole Pit areas. Additional Rotliegend fields occur in the Vlieland Basin and the north-western sector of the Central Netherlands Basin (Fig. 12).

In the north-east Netherlands, the giant Groningen field dominates (Roels, 2001; Verberg, 2001; Fig. 15). Other

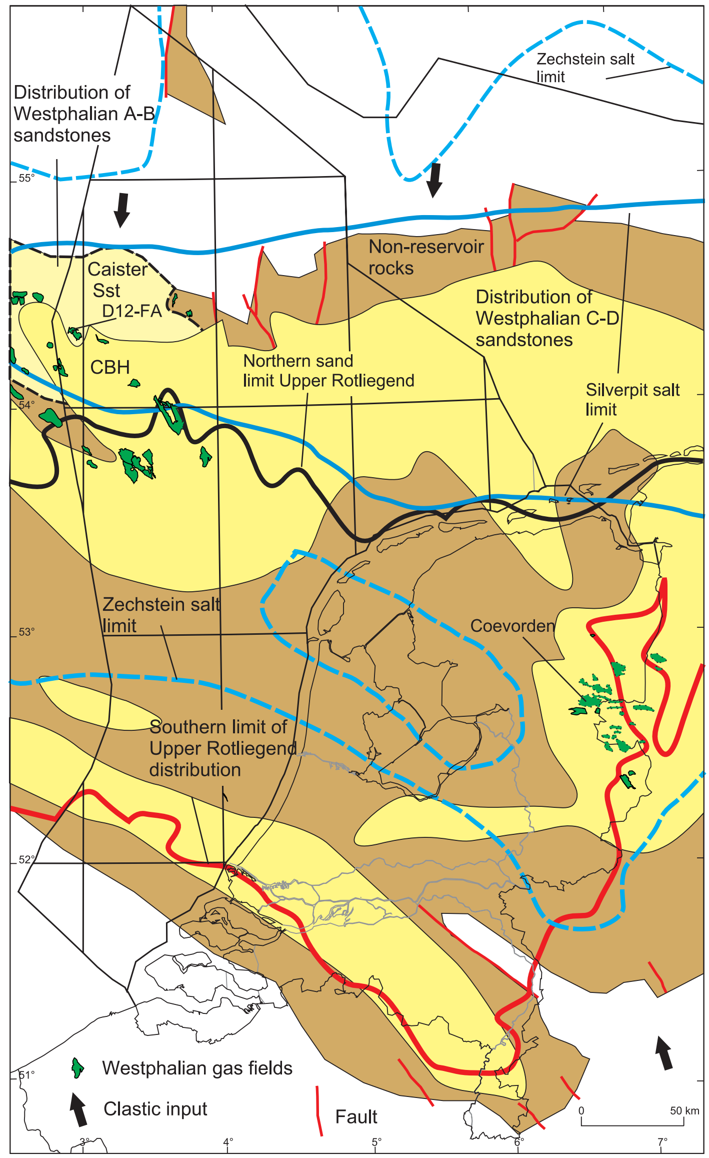


Fig. 10. Westphalian play map. This map outlines the distribution of the Westphalian reservoir rocks. The best reservoirs occur mostly in the sand-prone Westphalian C and D (yellow). On the Cleaver Bank High (CBH), gas has also been discovered in Westphalian A and B reservoirs (Caister Sandstone, light yellow). The names of some of the main fields are indicated. Most gas fields producing from the Westphalian occur in traps where the Upper Rotliegend sandstones are thin or absent, and where Zechstein salt or Silverpit shales provide a seal. In traps where Rotliegend sandstones are present, the Westphalian contains gas only where the height of the trap exceeds the thickness of the Rotliegend, such as in the Groningen field (not indicated on the map).

large accumulations are Annerveen (Veenhof, 1996) and the Ameland complex of fields (Crouch et al., 1996), both with a gas-initially-in-place (GIIP) volume of ca. 75×10^9 m^3. Noteworthy is that the Ameland complex appears to be underfilled. Structural reconstruction shows that

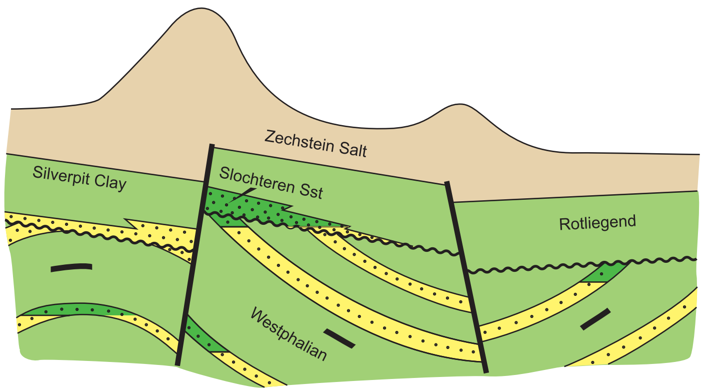
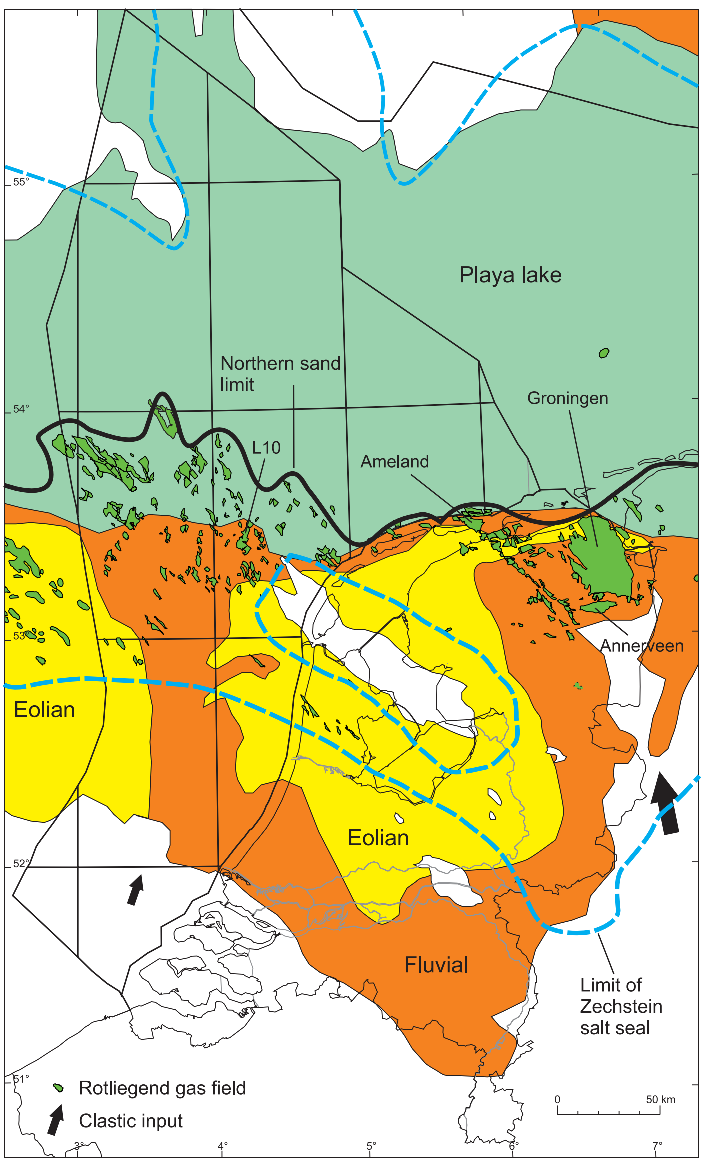


Fig. 11. Trap styles in the Westphalian. Most gas (dark green) is found directly below the Base Permian Unconformity, where Slochteren sandstones are thin or absent. At intra-Westphalian levels only sub-commercial gas volumes have been encountered to date.



Late Cretaceous inversion enlarged the trap. Apparently, subsequent charge was not sufficient to refill it.

Marked variations in the nitrogen contents in fields in the north-east Netherlands are attributed to variations in charge history. Traps that have access to recent charge are characterised by very low nitrogen contents (< 5%). On the Friesland Platform, where charge modelling indicates that the last phase of gas charge occurred prior to Jurassic rifting, nitrogen levels are as high as 25%. Residual oil is often reported from cores cut in Slochteren sandstones. This is interpreted to indicate early oil charge from Zechstein carbonate or Coppershale source rocks. The oil was later flushed out by excessive amounts of gas, filling the majority of structures to spill point. The only live oil found so far in Slochteren sandstones is in Midlaren in north-east Drenthe, where below a gas cap, a 120-m oil column was encountered, juxtaposed against Z2 Carbonate. It may be that the carbonates have allowed leakage of gas, but form a lateral seal to oil.

In the offshore K and L quadrants, K8-FA is the largest field with a GIIP of some 70×10^9 m^3. Traps here are also simple horst blocks overlain by a thick Zechstein evaporite as top seal. As the Ten Boer shales are much thinner than in the north-east Netherlands, there is less potential for additional column as a result of sealing Ten Boer in juxtaposition. Locally, where the Zechstein seal is breached by erosion, basal Triassic shales prove to be an effective top-seal (e.g. in K15-FC and K15-FK). Variations in gas-water contacts in different field compartments, and gas columns below conventional leak points, show that sealing faults are present also in the offshore. This has also been observed in the western continuation of the Rotliegend play in the UK sector of the southern North Sea (Leveille et al., 1997).

In the northern K blocks and adjacent areas, basal Slochteren sandstones, sealed by the overlying Silverpit Formation, contain commercial gas accumulations. The most significant is the Markham field in the J3/J6 area, which field straddles the UK/Dutch border. The basal Rotliegend sandstones show a greatly varying reservoir quality and thickness, hampering exploration efforts for this objective.

In the western sector of the Central Netherlands Basin, Rotliegend gas fields such as Bergen, Groet, Bergermeer, Schermer and Middelie have been brought to production (Van Lith, 1983).

Fig. 12. Upper Rotliegend play map. The gross depositional environments at approximately Upper Slochteren level are shown. The gas fields located in the area of playa-lake facies are in older, basal Rotliegend sandstone reservoirs. The southern limit of Rotliegend gas fields is controlled by the southern extent of the Zechstein top seal. The names of some of the main fields are indicated.

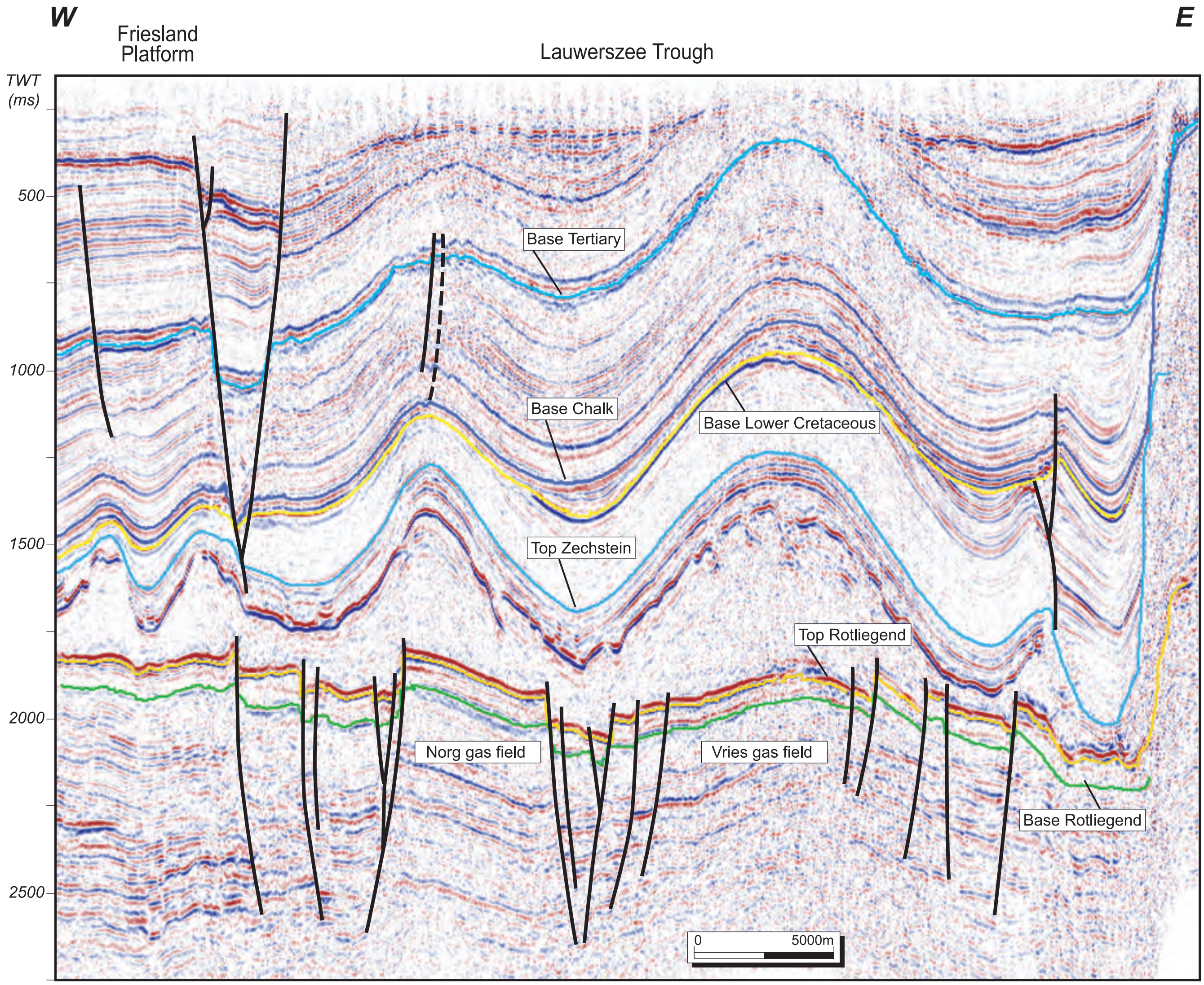


Fig. 13. Seismic line in the north-east Netherlands (Friesland Platform to Lauwerszee Trough) showing gas fields in fault blocks at Rotliegend level.

Even at significant depths, reasonable porosities (> 15% at 4000 m) occur in the better reservoir levels of the Slochteren. The main deterioration of reservoir potential is by early carbonate cementation or the growth of fibrous illite, which is associated with deep paleo-burial, in particular within the inverted Broad Fourteens Basin. Illite growth, which reduces permeability, increases where formation waters rich in ions of clay minerals are present, for example where the Slochteren is juxtaposed against the shaly Westphalian (Gaupp et al., 1993). However, where the Slochteren was already gas-bearing prior to deep burial, illite growth may have been prevented. Therefore, to be able to predict reservoir quality, a good understanding of depositional trends, as well as of timing of trap formation, gas charge and temperature history, is required. Low permeabilities and ensuing uneconomic production rates are also associated with depositional facies, such as fine-grained, argillaceous wadi sands (Frikken, 1999).

Modelling studies show that gas generation from the Westphalian coals in the central parts of the Broad Fourteens Basin peaked during Late Jurassic times, and ceased by the Late Cretaceous. This is consistent with K/Ar dates, which show that the latest illite formation in the gas-bearing reservoirs was generally in the Late Jurassic (Lee et al., 1985, 1989). Structurally valid, water-bearing traps within the basin are generally characterised by the presence of reverse faults, indicating that they were formed or modified during the Late Cretaceous inversion, i.e. after the main charge phase. On the margins of the basin, generation of gas continued to the present day.

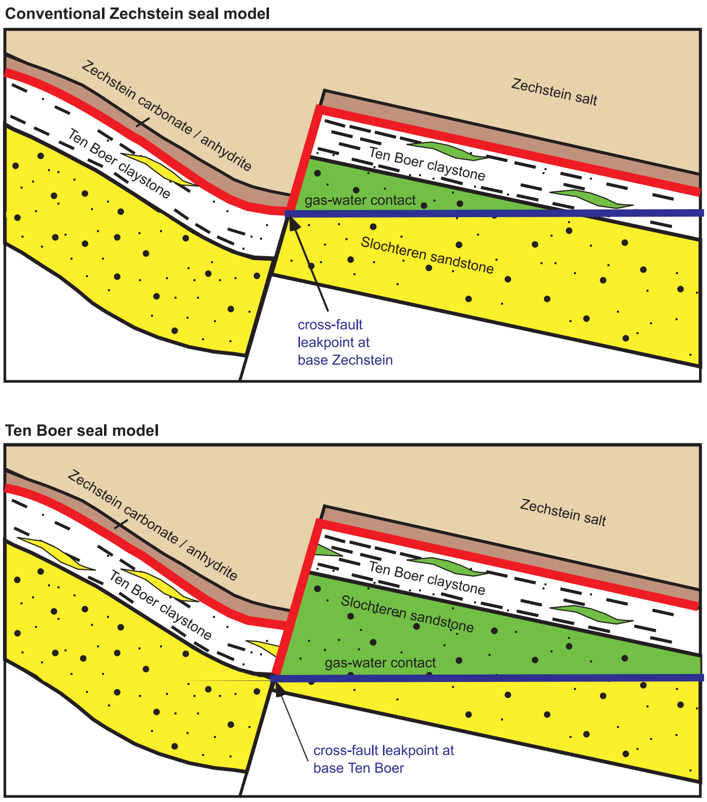


Fig. 14. Ten Boer sealing concept. The regional top seal for the Rotliegend is the base of the Zechstein (red line in upper image). Where the Ten Boer Member is sufficiently shaly, it provides a seal in fault juxtaposition (red line in lower image), even though intra-Ten Boer sandstones may be gas-bearing, and on the same pressure gradient as the gas in the Slochteren sandstones.

Zechstein play

Gas fields producing from the Zechstein occur in the Lower Saxony Basin in the east, and the Central Netherlands Basin in the west of the Netherlands (Fig. 16). The reservoirs are platform carbonates that developed along the southern margin of the Southern Permian Basin, and that are sealed by Zechstein salt and anhydrite (Van der Baan, 1990; Van de Sande et al., 1996; Geluk, this volume). The traps are predominantly fault-dip closures, with some four-way dip closures. Like for the Rotliegend play, charge comes from the Westphalian and Namurian. Contributions from Zechstein oil source rocks locally result in condensate-rich gas. Oil shows in the Zechstein have been recorded in numerous wells in the north-east Netherlands. Analyses indicate that they are derived from intra-Zechstein source-rock intervals and the Coppershale.

The best Zechstein reservoirs (porosities of ca. 14%) occur in the Z2 Carbonate and, less important, the Z3 Carbonate. A single accumulation in the Z1 Carbonate (Q5-A) came on stream late 2004 (Geluk, 2000; Ministry of Economic Affairs, 2006). The highest porosities are found in the shelf-edge facies that developed above the paleo-highs formed by the underlying Z1 Anhydrite. Sub-aerial exposure on the platform occurred occasionally and karstification locally enhanced the reservoir characteristics. The adjacent slope facies comprises redeposited platform sediments. Small-scale reefs occur locally, but they are restricted to the Z1 Carbonate (Geluk, this volume). The basinal facies of the carbonates comprises fine-grained limestones and dolomites. The Zechstein carbonates experienced a complex diagenetic history, including cementation, leaching and dolomitisation, which resulted sometimes in strong lateral variations in their porosity and permeability. Especially in the tight slope and basinal facies, with porosities of less than 4 or 5%, e.g. the Dalen, Emmen and Coevorden fields, economic gas production rates depend to a large extent on the presence of fractures (Frikken, 1999).

Zechstein fields also occur in the west of the Central Netherlands Basin, where the Alkmaar field is now used for underground gas storage (Bos, this volume).

The Zechstein carbonate platforms along the southern border of the Southern Permian Basin continue into the Dutch offshore. The first offshore gas discovery, in 1968, was actually made in Zechstein dolomites in block P6 (Van der Poel, 1989). Various additional wells in the P and Q quadrants and the southernmost part of the K quadrant have subsequently tested gas, but only in block P6 is gas being produced from the Zechstein below a larger Triassic gas accumulation. Along the northern margin of the Southern Permian Basin, in the area of the Elbow Spit and Mid North Sea highs, the Zechstein is also developed in slope and platform facies as proven by well E2-2, but is of only limited thickness. Towards the north, its reservoir quality may improve.

A special case is the gas accumulation discovered in 1985 by well G16-1 on the Schill Grund High, where a leached Z3 Carbonate and an overlying Upper Jurassic sandstone form a single composite reservoir above a Zechstein salt dome. The Lower Cretaceous Vlieland Claystone Formation provides the seal.

In the West Netherlands Basin, the Zechstein is partly represented by a siliciclastic basin-margin facies. The best reservoir development occurs offshore where, in block P10, a sandstone interval reaches a thickness of ca. 125 m. However, no Zechstein evaporites are present here to provide a top seal, and only sub-commercial quantities of gas have been encountered in the area.

Triassic plays

Triassic sandstones are widely distributed in the Netherlands and form, after the Rotliegend, the second most important gas reservoirs. Producing fields occur in the east Netherlands, in the West Netherlands Basin, and in the northern and western offshore (Fig. 17). The top seal is

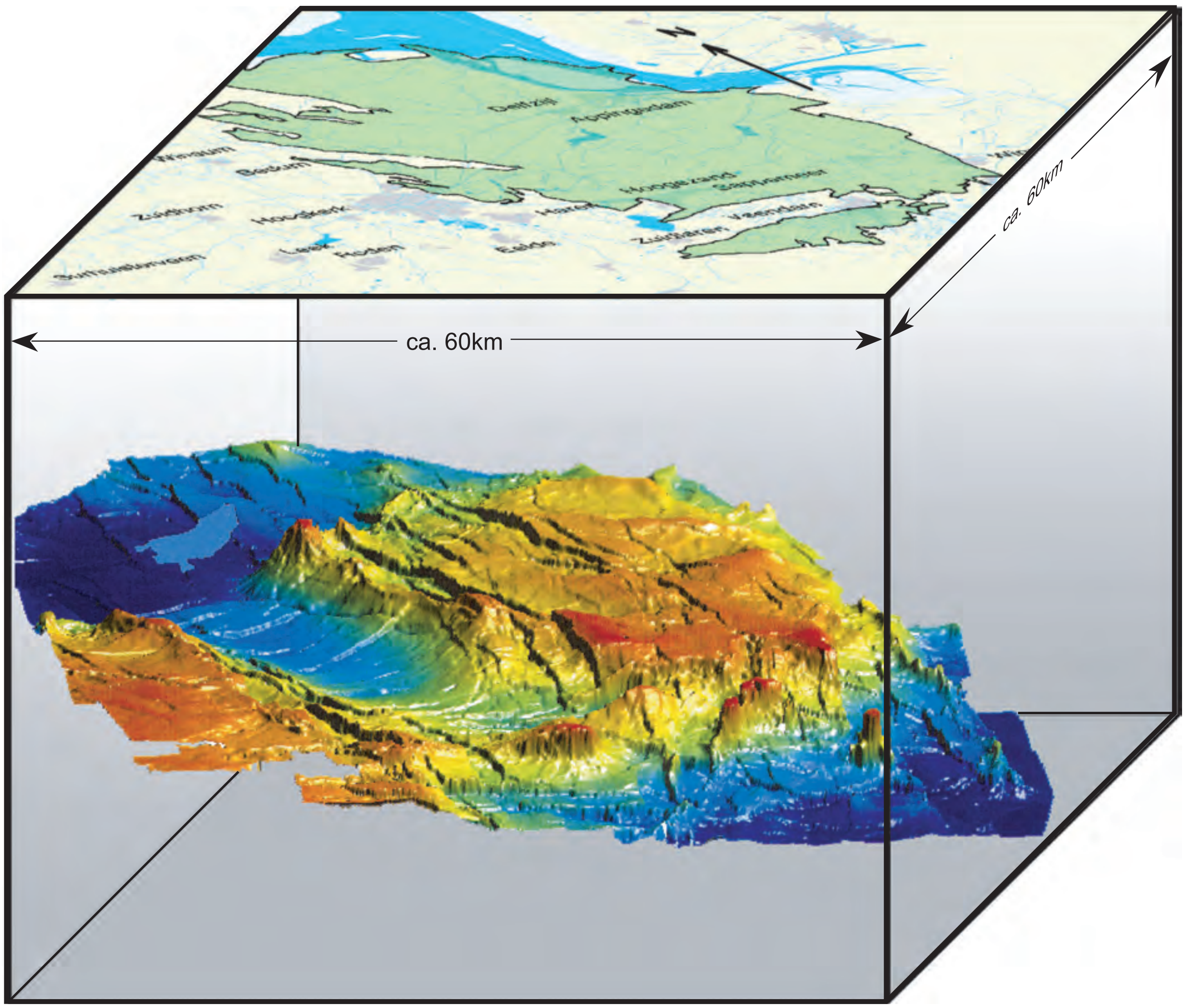


Fig. 15. 3D view of the Groningen field, with an initial recoverable gas volume of ca. 2700 × 10^9 m^3, the largest gas field of Europe. The level shown is the top of the Upper Rotliegend Group. To the west of the Groningen field (note the north arrow), the depression of the Lauwerszee Trough shows up in blue colours, with the Friesland Platform further to the west in yellow to reddish colours.

formed in most cases by Triassic shales, and where the Triassic is truncated at the Base Cretaceous Unconformity, by Lower Cretaceous shales. The gas comes mainly from Westphalian source rocks and occurs in a variety of trap styles. Where the Zechstein salt is present, as in the northern sector of the Dutch subsurface, it forms a barrier preventing Westphalian gas from reaching the Triassic reservoirs. Only where this regional seal is breached, by salt withdrawal or faulting, can Westphalian gas reach the Triassic reservoirs. In the West Netherlands Basin no Zechstein salt is present, and there the Triassic sandstones form the first well-sealed reservoir above the Westphalian.

The Triassic sandstones were deposited in fluvial and eolian settings. They were derived from the south, and to the north progressively more shales are present. In the south, the Main Buntsandstein forms a thick massive reservoir package. To the north this package breaks up into several thinner sandstone units, such as the Volpriehausen and Detfurth sandstones, which are separated by clay- and siltstones that form regional seals (Ames & Farfan, 1996; Geluk, 1999).

In the West Netherlands Basin, the exploration for gas in the Triassic really started only in 1982. Currently nine gas fields are producing, and several others remain undeveloped. The typical trap in this play consists of a Late Jurassic horst block in which the reservoir is sealed vertically by Upper Triassic evaporitic shales, and laterally by Upper Triassic to Lower Jurassic shales (Fig. 18; De Jager et al., 1996). Lateral seal risks exist where fault-throws are so large that cross-fault juxtaposition is against the sandy Delfland sequence. The best reservoirs occur along the south-western basin margin, where excellent gas produc-

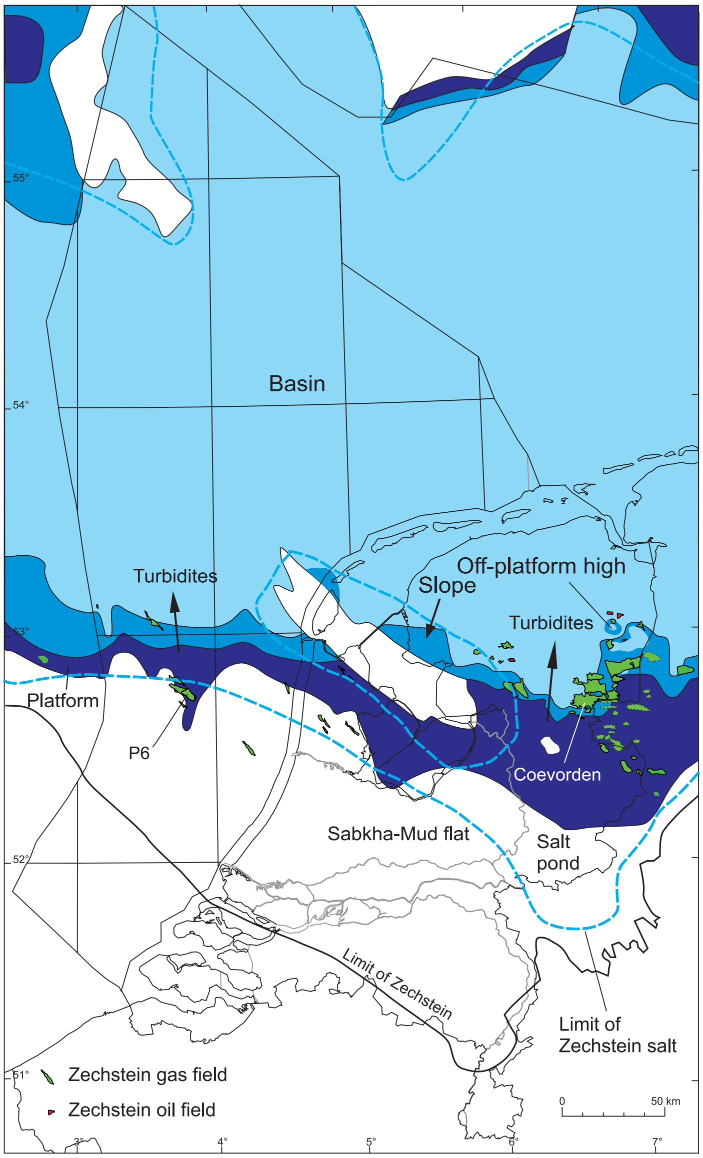


Fig. 16. Zechstein play map. Shown is the paleogeographic map of the Z2 Carbonate, which constitutes the main Zechstein reservoir where it is in platform facies. Gas accumulations in the western offshore, located south of the distribution area of this carbonate, are in the Z3 Carbonate. The location of the Coevorden field is indicated.

tion rates of several million cubic metres per day have been achieved. Towards the north, the reservoirs rapidly deteriorate because of erosion of the best reservoir section in the upper parts of the Main Buntsandstein and a general increase in pre-inversion burial, which caused diagenetic reduction of porosity and permeability. Furthermore, as a result of tectonic inversion during the Late Cretaceous, traps in the central and northern part of the basin changed their configuration or were destroyed, while new ones formed. At the same time, both the Carboniferous and the Jurassic source rocks largely stopped generating hydrocarbons in this part of the basin following uplift and declining heat flow (De Jager et al., 1996). A possible timing problem thus exists for new traps that formed during the inversion. In the southern part of the West Netherlands Basin, where no or very little inversion occurred, gas generation continued to the present day. Although the Triassic play in the basin is largely a gas play, in several accumulations (Papekop, Pernis West, Spijkenisse Oost, Waalwijk Zuid and Botlek) an oil leg was found underneath the gas, and the Ottoland structure probably contains oil only. The oil has been correlated to the source rocks of the Lower Jurassic Posidonia Shale, with probably some contribution from the Lower Jurassic Aalburg and uppermost Triassic Sleen formations. This oil migrated into the older reservoirs from down-thrown blocks across faults. Where the Posidonia Shale has not been down-faulted to below the Triassic reservoir, as is the case for some oil-bearing structures, it is very unlikely that it could have charged these traps. There the oil is most probably derived entirely from the Aalburg and Sleen formations (Van Balen et al., 2000).

In the western sector of the Central Netherlands Basin, the Middelie, Bergen and Q8 fields also contain gas in Triassic sandstone reservoirs.

In the northern offshore, the Volpriehausen and Detfurth sandstones are reservoirs for gas in the northern Vlieland Basin, Terschelling Basin, Schill Grund High and southern Dutch Central Graben area. In 1992, the L5-FA field was the first main Triassic field in the northern offshore to come on stream. Further economic accumulations are known in F15 (Fontaine et al., 1993), G17 and L2. In these blocks most traps are related to halokinesis of the underlying Zechstein salt, and occur in turtleback anticlines, against salt walls or as fault-dip closures (Fig. 19). Salt walls in the area of the Terschelling Basin and southern Dutch Central Graben seal off separate pressure cells in which the Triassic reservoirs occur (Crepieux et al., 1998). In some of these cells reservoir pressures equal the minimum horizontal stress, thus causing seal breach. Deeper culminations in these pressure cells may be protected against seal failure.

Sandstones of the Triassic Solling Formation present a unique reservoir locally. Well L9-8, drilled in 1993, found this reservoir, which was hitherto unknown, gas-bearing. Subsequent development drilling of the L9-FD, FF and FI complex of fields established total recoverable gas reserves of ca. 26×10^9 m^3 in the reservoir which is up to 125 m thick and has a net over gross ratio of close to 100%. Gas production rates from this extremely prolific reservoir, with average permeabilities in the Darcy range, are typically several million cubic metres per day per well. To the north, in block L6, the same reservoir is salt-plugged. The great thickness of the reservoir in part of block L9 suggests a unique depositional setting. Extensive core coverage indicates that the reservoir is made up mainly of eolian sandstones, with locally some water-laid sediments. Well and seismic data show that the reservoir package is wedge-shaped, thickening into a listric normal fault that detaches onto the top of the Zechstein salt, adjacent to a salt wall. The eolian sandstones were clearly preserved in a syn-depositional half-graben, the development of which appears to be related to Early Triassic extension.

Salt plugging of the reservoirs is a serious risk to the Triassic play in the northern offshore (Purvis & Okkerman, 1996). Salt-plugged reservoirs have been encountered in particular near salt walls and along fault planes, and are often characterised by a phase reversal of the seismic response of top reservoir. Careful study of seismic amplitudes may indicate whether good reservoirs are present or not.

On the Cleaver Bank High, no discoveries have been made in the Triassic so far. Because of deep erosion, the Triassic is in many places truncated at the Base Cretaceous Unconformity. Associated leaching has resulted in excellent reservoirs, with average porosities of up to 27%. The lack of exploration success is attributed to a lack of charge windows through the thick sealing claystones and evaporites of the Rotliegend and Zechstein.

In the western offshore the P6 fields produce gas from Triassic sandstones. The Buntsandstein is also a proven gas reservoir in the eastern Netherlands. The Wanneperveen and De Wijk fields (Bruijn, 1996) produce from reservoirs that are sealed by shales of the Lower Cretaceous Vlieland Claystone Formation. Reservoir characteristics are enhanced as a result of leaching below the Base Cretaceous Unconformity. Even leached anhydritic shales of the Lower Triassic Rogenstein are productive in these fields. Base Vlieland channel sands locally form part of the complex traps.

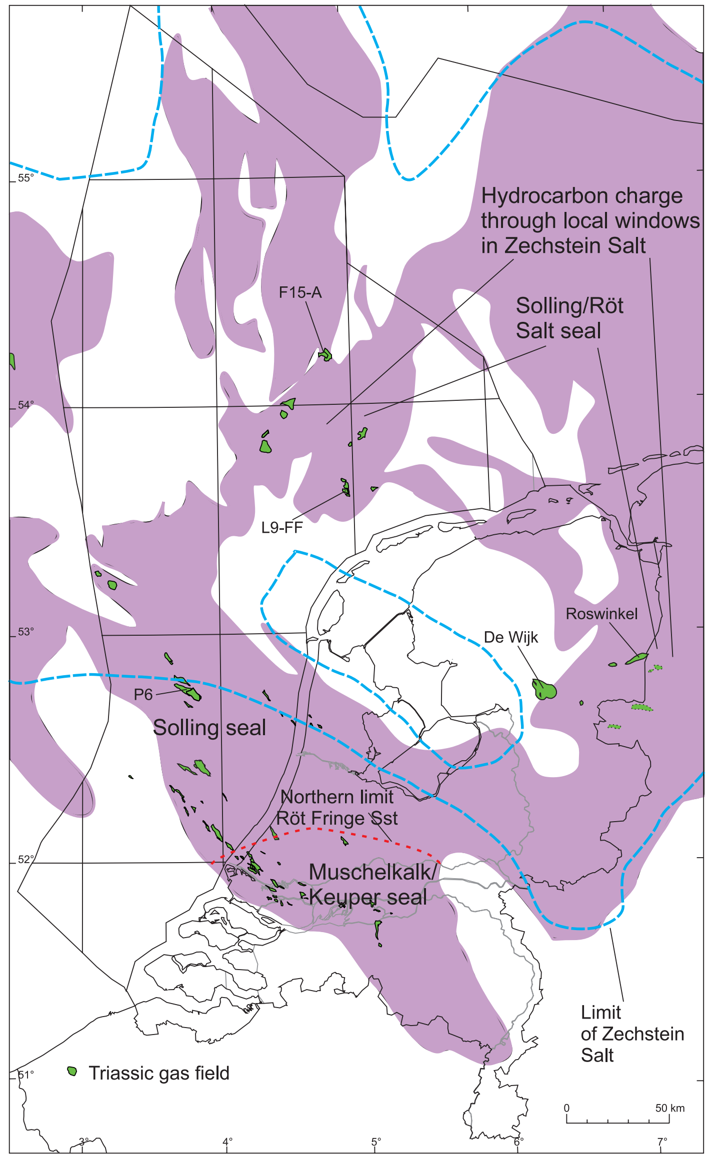


Fig. 17. Triassic play map showing gas fields producing from Triassic reservoirs and the present-day distribution of the Main Buntsandstein Subgroup (purple) and the Röt Fringe Sandstone. These units form the main reservoirs. The main top seals are indicated per area. Beyond the depositional limit of the Zechstein salt, Westphalian gas can migrate directly into Triassic reservoirs. Where the Zechstein salt is present, charge into the Triassic relies on the presence of breaches in the Zechstein as a result of salt withdrawal or faulting. The names of some of the main Triassic fields are indicated.

Upper Jurassic and Lower Cretaceous plays

The Upper Jurassic and Lower Cretaceous oil plays are restricted to the Kimmerian rift basins, such as the Dutch Central Graben and the Broad Fourteens, West and Central Netherlands, and Lower Saxony basins. The reservoirs occur in clastic syn-rift and early post-rift deposits, which show rapid facies variations, ranging from continental to marine. There is also a great variation in trap styles. These plays are much more oil-rich than the other Dutch plays. This is mainly because of charge from the Lower Jurassic Posidonia Shale, which has been preserved only within the extensional Late Jurassic-Early Cretaceous rift basins (Fig. 9). Most of the gas present is derived from the Westphalian, and in the Dutch Central Graben also from Jurassic coal-bearing sequences and/or deeper Jurassic shales of the Aalburg Formation.

In the east Netherlands, the Schoonebeek field has produced a total of ca. 250 million barrels (40×10^6 m^3) of oil from the Lower Cretaceous Bentheim Sandstone, at a depth of ca. 800 m. The accumulation is rather exceptional, as it is sourced from lacustrine algal Type-I source rocks of Early Cretaceous age (Coevorden Fm). These source rocks have a limited distribution in the Netherlands, but extend into the Lower Saxony Basin in Germany (Binot et al. 1991). The relatively low gravity of the oil of

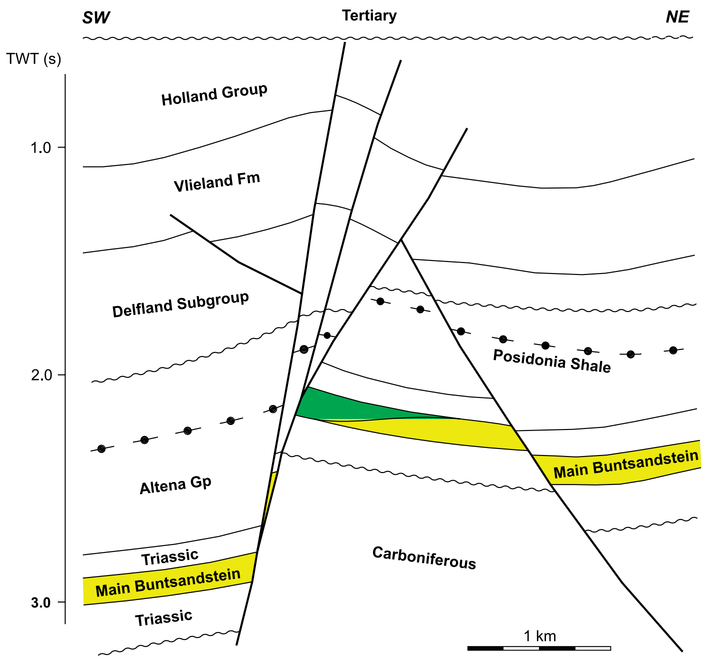


Fig. 18. Example of a trap in the Triassic of the West Netherlands Basin (Wassenaar Deep gas field). The Main Bundsandstein reservoirs (yellow) have younger Triassic shales as top seal, and are laterally sealed by Lower Jurassic shales.

25° API results from the low maturity of the source rocks. Geochemical data indicate that the oil is an early expulsion product, without any sign of biodegradation.

Gas fields at the level of the Lower Cretaceous Vlieland sandstones in the north-east Netherlands are, amongst others, Tietjerksteradeel, Wanneperveen, Leeuwarden, Harlingen (Van den Bosch, 1983) and De Wijk. Additional discoveries have been made in the Lemmer-Steenwijk concession. The main difficulty of this play is to predict the presence of the Vlieland sandstones, the distribution of which seems somewhat erratic and cannot readily be interpreted from the seismic data.

In the West Netherlands Basin, an intensive drilling programme in the 1950s led to the discovery of the Rijswijk, Pijnacker, De Lier, IJsselmonde, Wassenaar, Zoetermeer and Moerkapelle oil fields. Exploration from the late 1970s to early 1990s resulted in further discoveries, both onshore (Berkel, Barendrecht, Rotterdam, Pernis and Pernis West) and offshore (P8a, P9, P11, P12, P15, and Q13). Rotterdam is the largest field, with initial reserves of ca. 90 million barrels (14×10^6 m^3). Both gas and oil are present in sandstones of the Upper Jurassic to Lower Cretaceous syn-rift and Lower Cretaceous post-rift deposits (Bodenhausen & Ott, 1981). Many of the older fields have been abandoned or closed-in over the last few years, and all current oil production is from Berkel, Rotterdam and Pernis West. The best reservoirs are the post-rift Rijswijk, Berkel and IJsselmonde sandstones that have been deposited as coastal barrier complexes overlying the Late Kimmerian Unconformity (Den Hartog Jager, 1996). Additional reservoirs are the younger De Lier and Holland Greensand sandstones and the older syn-rift Delfland Subgroup. Many of the reservoirs have oil columns with a gas cap, with the younger reservoirs containing generally more gas (De Jager et al., 1996). The presence of gas is often a downgrading factor, as it reduces the potential oil volumes in traps, while it is not sufficient to justify development as gas fields.

Some of the fields contain very heavy oil with API gravities of 13 to 20°, resulting in low recovery factors (e.g. Moerkapelle and Wassenaar). This is the result of biodegradation at the onset of the Tertiary. Because of tectonic inversion, reservoirs were then at or near the surface, where bacteria had access to fresh meteoric waters (De Jager et al., 1996).

Traps occur mainly as four-way dip-closures along anticlinal trends that formed in response to the Late Cre-

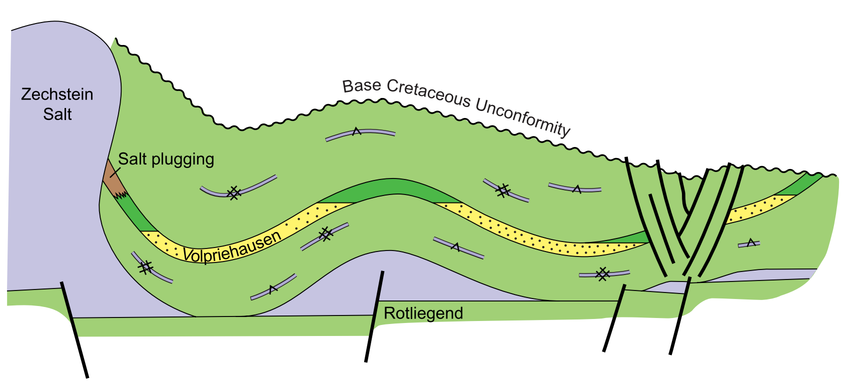


Fig. 19. Trapping styles in the Triassic of the northern offshore. Most gas (dark green) is trapped in four-way dip-closed structures above salt swells or in turtle-back structures. Where the Triassic is truncated at the Base Cretaceous Unconformity, Lower Cretaceous marine shales may provide the seal. Salt-plugged reservoirs may provide lateral stratigraphic seals.

taceous and Early Tertiary inversion along the south-western basin margin (Fig. 20; Racero-Baena & Drake, 1996). Further north, in the more strongly inverted sectors of the West Netherlands Basin, and also in the Central Netherlands Basin, the reservoir objectives have been eroded. Although the rapid facies variations within the Upper Jurassic and Lower Cretaceous suggest a potential for stratigraphic traps, these have yet to be encountered.

The Upper Jurassic and Lower Cretaceous oil and gas plays of the Broad Fourteens Basin are essentially the same as in the West Netherlands Basin, with the same source rocks and reservoir sandstones. All known trapped hydrocarbons are located along the north-eastern margin of the inverted basin in Q1 (Helm, Hoorn, Helder, Haven and Halfweg fields; Roelofsen & De Boer, 1991) and in K18 and L16 (Kotter and Logger fields; De Jong & Laker, 1992; Goh, 1994). In the inverted areas of the Broad Fourteens Basin, as in the West Netherlands Basin, there is a timing problem with regard to hydrocarbon generation and trap formation. In these areas, the Carboniferous source rocks became mature in the Late Jurassic or the Cretaceous, prior to the main phase of trap formation, which is Late Cretaceous or even younger.

In the Dutch Central Graben, several oil discoveries have been made in Upper Jurassic and Lower Cretaceous sandstones (Wong et al., 1989; Wong, 1991). The largest field is F3-FB, where several sandstones in the Upper Jurassic Central Graben Group contain light oil (ca. 55° API) and gas. The volume of oil initially in place in this field is ca. 100 million barrels (16×10^6 m^3). Other oil accumulations are F3-FA, F14-FA, F17-FA, F17-FB, F18-FA, L1-FB, L2-FA and L5-FA. Gas caps are present in F3-FB and F18-FA, while in F3-FA mainly gas is trapped. Most of these hydrocarbons have been generated by the Posidonia Shale, with minor contributions from the Aalburg and Sleen formations and from coals in the Central Graben Group.

The traps in the graben occur as four-way dip-closures in turtle-back structures (F3-FB, Fig. 21) and as tilted fault blocks. The southern sector of the graben has been more strongly inverted, resulting in greater structural complexity and compartmentalisation of the traps. Reservoir distribution is another important factor in the syn-tectonic sequences that have been deposited in fluvial, deltaic and lagoonal environments. This is in particular the case in the south of the graben, where there are great variations in sand to shale ratios, and where individual sands seem to have a limited lateral extent.

In the Terschelling Basin only one small gas discovery has been made in block L6. This lack of success is attributed mainly to a lack of lateral and top seals. Gas charge may be limited as well, as migration paths from the Westphalian through the Zechstein evaporites are dif-

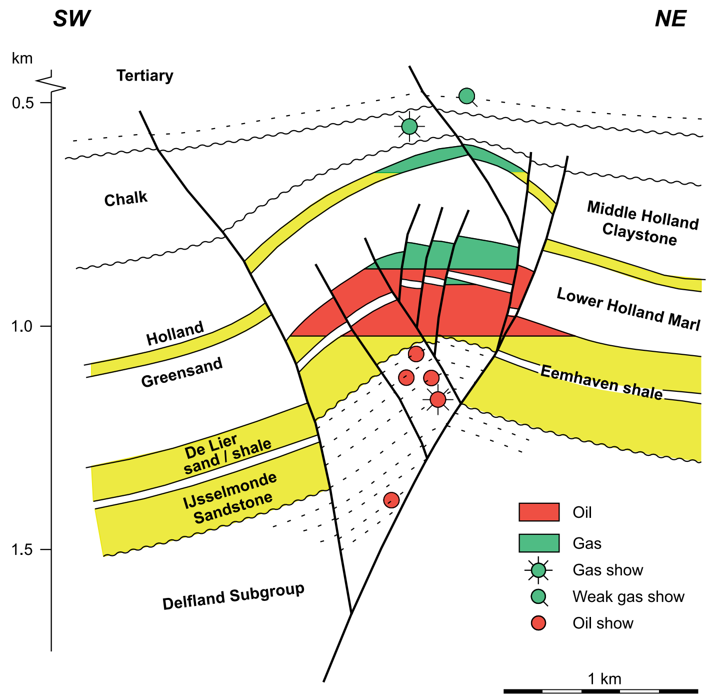


Fig. 20. Example of a trap in the Upper Jurassic and Lower Cretaceous in the West Netherlands Basin (IJsselmonde-Ridderkerk field). Several stacked reservoirs are present, the younger of which contain more gas and the deeper more oil.

ficult to establish. Coals within the Central Graben Group are not very extensive, but may have contributed some gas charge. In the Waddenzee area of the Vlieland Basin, the Zuidwal field produces gas from the Vlieland Sandstone Formation in a drape structure over an Upper Jurassic volcanic complex (Perrot & Van der Poel, 1987; Herngreen et al., 1991).

Upper Cretaceous Chalk play

The Chalk Group is the reservoir for several major oil fields in the Danish sector of the North Sea (e.g. Dan, Gorm and Skjold). In the Netherlands the first and so far only economic oil accumulation offshore is the Hanze field, discovered in 1996 in block F2 in the north of the Dutch Central Graben. In several structurally valid tests, oil shows have been encountered in the chalk, in particular in the Dutch Central Graben area. In most of these cases the oil is likely to have been derived from the Posidonia Shale.

The only onshore gas field producing from the Chalk Group is the Harlingen field (Van den Bosch, 1983). Gas is also present in the upper part of the Chalk Group at De Wijk and Wanneperveen, where the gas columns extend upwards into the Tertiary Basal Dongen Tuffite. Anticipated subsidence problems prevent development of these reserves. In the West Netherlands Basin, gas has been

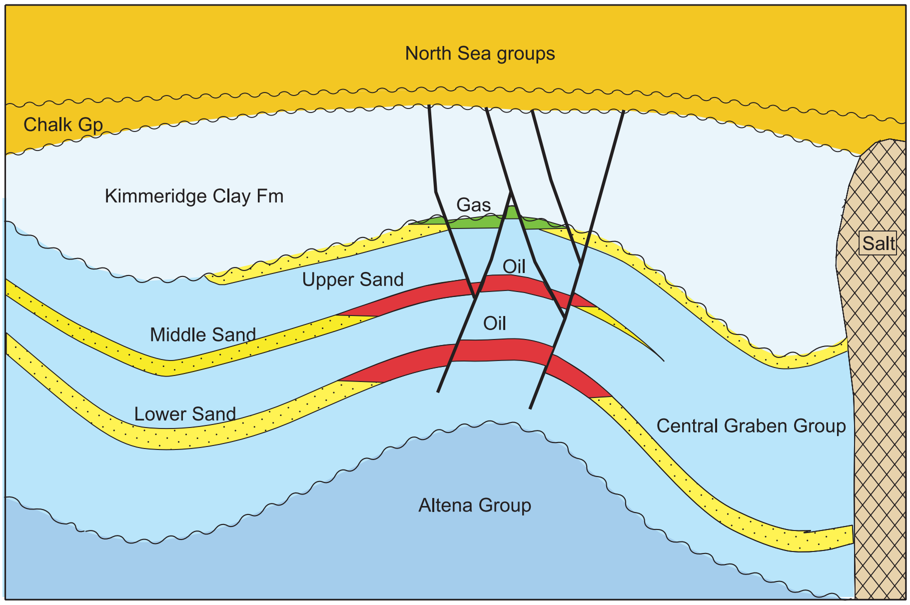


Fig. 21. The F3-FB field in the north of the Dutch Central Graben. Three main Upper Jurassic sandstone reservoirs are present in a faulted four-way dip-closed turtle-back structure. The reservoirs are not in communication and have separate hydrocarbon-water contacts.

tested from the Maastrichtian and Danian parts of the Chalk Group in the IJsselmonde structure.

Although the reservoir characteristics of the Chalk Group are generally only poor to fair, they are comparable to those of the chalk in the Danish and Norwegian sectors of the North Sea. The lack of success in the Chalk in the Netherlands is possibly related to the relatively shallow depth of burial and to a limited charge. The shallow burial has resulted in a reduced sealing capacity of the Lower Tertiary above the somewhat overpressured chalk reservoirs, causing leaky traps.

Tertiary and Quaternary plays

In the IJsselmonde structure in the West Netherlands Basin, small amounts of gas have been tested from the Lower Tertiary Basal Dongen Sand, which lies some 30 m above the base of the Lower North Sea Group, and has average porosities ranging from 34 to 39%. In the east Netherlands, the gas trapped in the Basal Dongen Tuffite in the De Wijk field has not been developed because of anticipated subsidence problems during production.

The main Tertiary-Quaternary gas accumulations occur in the A and B blocks of the northern offshore at depths of 400 to 700 m. Strong amplitude anomalies and deeper pull-downs on seismic profiles clearly indicate the presence of gas in subtle structures of only several tens of metres height. Most of the shallow gas accumulations are associated with gas chimneys, indicating leakage. Although the gas-bearing sandstones had been seen on seismic, and were encountered in wells with deeper objectives, it took until 1988 before well A12-3 tested potentially economic production rates from the shallow gas discovered by well A12-1 in silty sands in the topsets of Plio-Pleistocene prograding shelf sequences. Further shallow gas accumulations were discovered in B10, B13 and B16-1. Sand production from these unconsolidated reservoirs poses a major development problem.

Future potential

Proven plays

The Netherlands must qualify as a mature hydrocarbon province. With most of the current licences covered with 3D seismic data, and with many wells drilled, the main prospective fairways have already been well explored. Yet, new discoveries continue to be made, and a plot of cumulative volumes of gas found over time does not show indications of a reduced exploration efficiency or 'creaming' (Fig. 22). The continued steady pace of new discoveries over the last two decades certainly results in part from the application of seismic 3D technology, which has revealed traps that hitherto went unnoticed. Related advances in seismic processing, such as pre-stack depth migration, have more recently further improved the accuracy of predictive subsurface models. In addition, geological concepts continue to be improved, resulting in the identification of new play opportunities and trapping styles. The future potential for new gas discoveries in the Netherlands, based on identified prospects in proven plays, as reported by the Ministry of Economic Affairs (2006), is esti-

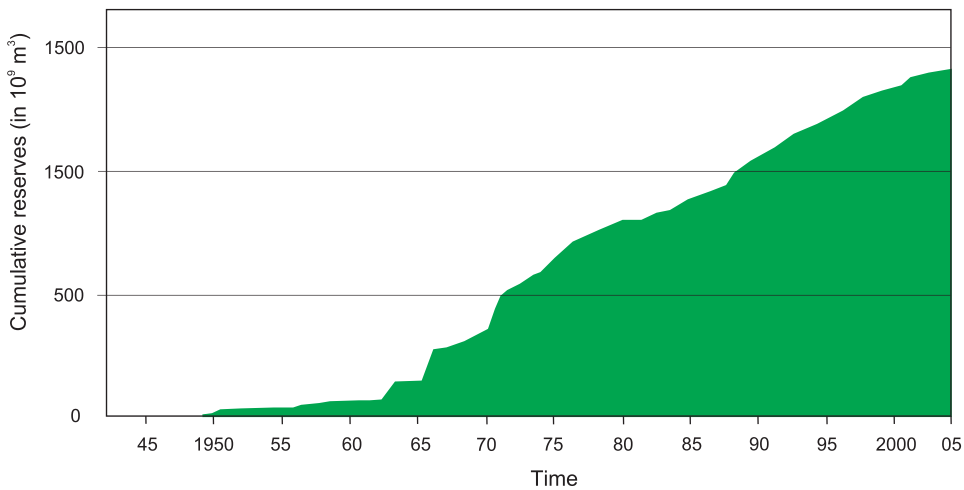


Fig. 22. Gas creaming curve for the Netherlands, excluding the giant Groningen field. Since the discovery of this field in 1959, an almost constant volume of new gas per year has been discovered. The flattening of the curve since 1998 is due to a reduced drilling activity.

mated to be between 180 and 440 × 10^9 m^3 of gas. These future reserves represent the addition of risked volumes of identified prospects in proven plays above a cut-off volume of 0.5 × 10^9 m^3 of gas onshore and 2.0 × 10^9 m^3 offshore. These numbers do not include the unidentified potential that may result from identification of new plays and prospects, for example through the application of new technologies or extensions to proven play areas.

In a densely explored area such as the Netherlands, ongoing exploration continues to lead to new insights. The great variety of structural styles and potential reservoir-seal pairs, only improves the chances for unidentified plays and prospects to occur. New technological developments are likely to provide increased accuracy to the predictions and a better understanding of play and prospect risks. A break-through in the area of improved production of gas from tight reservoirs may not only allow the exploitation of many accumulations that are currently uneconomic, but will also broaden the scope for exploration. Similarly, a break-through in the area of permeability prediction (e.g. from seismic data) would result in the derisking of prospects that are currently highly risky. And, last but not least, the lowering of economic thresholds or cheaper drilling and production technology may render presently non-commercial accumulations economic in the future.

New or speculative plays

In most of the Netherlands the Devonian occurs deeper than 5 or 6 km, and is a speculative objective at best. Dinantian carbonates on the northern flank of the London-Brabant Massif and in well Winterswijk-1 in the east Netherlands have been found tight and water-bearing. Nevertheless, some potential remains in karstified and fractured platform-edge carbonates sealed by Namurian shales that may contain source rocks. Wells have indicated good porosities and permeabilities (Darcy scale) in northern Belgium, where gas is stored in Dinantian rocks near Loenhout, east of Antwerp.

In the northern offshore, the mixed clastic and carbonate sequences of the Dinantian and Namurian Yoredale Formation, sealed by intraformational marine shales, form a speculative play. Charge could come from Visean coals or marine Namurian shales. There is a high risk, however, that these potential source rocks are post-mature.

Intra-Westphalian gas accumulations have to rely on intra-formational seals and on lateral seals by favourable juxtaposition across faults. That this may work is proven by a small and uneconomic gas discovery in block Q13. Also elsewhere in the Netherlands, intra-Westphalian shales occasionally seal small gas columns. Although no regional Westphalian seals have been identified, the 'marine bands', which form laterally extensive but thin shale intervals, could prove to be effective as intra-formational seals.

Gas accumulations may be present in Rotliegend sandstones along the northern margin of the Southern Permian Basin in the A and B quadrants. Prospectivity in this area relies on finding structurally valid and well-sealed traps. A risk is the absence of Westphalian source rocks. However, Namurian shales and possibly Visean coals display source-rock quality in the nearby German and British wells.

Elsewhere, remaining Rotliegend potential is likely to be present in down-faulted traps relying on sealing faults. A technological break-through in the production of gas from tight reservoirs (permeability < 1 mD) would open up additional potential.

Two potential Zechstein hydrocarbon objectives have been proposed by Geluk (2000). Firstly, Z1 reefs, which could be present along the Dutch Central Graben and the Mid North Sea High and in the eastern Netherlands, as well as above topographic base-Zechstein highs elsewhere. Secondly, Z2 carbonates in slumps and mass flows in the western offshore, in front of a carbonate platform that existed in the southern K and L quadrants. Furthermore, Z2 off-platform highs may occur in this area, even though they have not yet been identified in wells. For successful exploration of these plays, 3D seismic interpretation and high-resolution sequence stratigraphy form a prerequisite.

Above the Zechstein salt, the Triassic is in many places intricately structured, and several potential reservoir objectives are present. The discovery in 1993, with the 8^{th} well in the 20 × 20 km L9 block, of a locally developed but exceptionally prolific sandstone package of up to 125 m thick in the Solling Formation suggests that other unknown Triassic reservoirs may be present as well. Unidentified trapping geometries may also occur in these sequences. The complex structuration, however, also implies a risk of compartmentalisation of potential gas accumulations.

Several oil discoveries in the Upper Jurassic and Lower Cretaceous of the Dutch Central Graben are sub-economic because of compartmentalisation related to the presence of sealing faults and to a complex depositional reservoir architecture. Improved understanding of their depositional setting may lead to a better assessment of risks and to improved prediction of reservoirs, thus creating possibilities for development of these accumulations.

Conclusions

Gas is by far the most important hydrocarbon resource in the Netherlands. The discovery of the giant Groningen field in 1959 triggered an exploration effort that has continued to the present day. As a result, ca. 56% of the Dutch on-and offshore areas are now covered by good-quality 3D seismic data, and just over 1500 exploration and appraisal wells have been drilled. By the beginning of 2006, a total of 4349 × 10^9 m^3 of gas had been found, of which 1510 × 10^9 m^3 remain to be produced. The Groningen field accounts for two thirds of the total Dutch initial gas reserves. The production of ca. 73 × 10^9 m^3 of gas per year nowadays provides the Dutch State with an annual income of ca. €7.5 billion.

Almost all the gas in the Netherlands has been generated from Westphalian coals. The main reservoir is the directly overlying Upper Permian Rotliegend sandstones, sealed by the thick Zechstein salt. The second objective for gas is the Triassic.

Most oil is generated from the Lower Jurassic Posidonia Shale. As this source rock is preserved only within Mesozoic extensional basins, most oil fields are found there. The total initial producible oil reserves are relatively modest and amount to just over 1 billion barrels (165 × 10^6 m^3), of which 234 million barrels (36 × 10^6 m^3) remain early 2006.

Additional discoveries continue to be made, with no indications of a reduction in exploration efficiency. The Ministry of Economic Affairs estimates the remaining undiscovered reserves in mapped prospects within proven plays to be between 180 and 440 × 10^9 m^3 of recoverable gas. The volumes associated with yet to be identified prospects and plays, and those that will result from improved technologies, will undoubtedly increase this remaining potential.

Acknowledgements

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Peat, coal and coalbed methane

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ABSTRACT

Several stratigraphic intervals in the subsurface of the Netherlands contain layers of peat or coal. These include Holocene peat, Miocene brown coal, and Upper Carboniferous (Westphalian) bituminous coal and anthracite, which have all played important roles as energy sources in the country. Imported coal still does. Total peat production throughout the centuries reached ca. 9.4×10^9 m^3, while peak annual brown-coal production reached 1.5 Mt. Total hard-coal production in the 20th century amounted to 568 Mt. Production of these energy sources ceased during the 20th century. Current resources of brown coal are estimated to be at least 1700 Mt; estimates for bituminous coal resources shallower than 1500 m are between 4000 and 38 000 Mt. Since the 1990s, interest has increased in coalbed methane, i.e. gas that is contained in bituminous coal. Production of coal or coalbed methane from the Dutch subsurface is currently not economically feasible. Estimates for the theoretically recoverable volumes of coalbed methane shallower than 1500 m are between 7 and 107×10^9 m^3. However, as a result of international measures to comply with the Kyoto protocol of 1997, coalbed-methane production might become economically feasible in combination with subsurface storage of CO_2. In this case, the coal is considered as a sink for the CO_2. The storage potential of coal seams shallower than 1500 m is estimated to be between 39 and 594 Mt of CO_2. Consequently, coal may still play a role in future energy supply as conventional fossil fuel resources decline and technology advances.

Keywords: Netherlands, brown coal, CO_2 sequestration, resources, Westphalian

Introduction

The world has relied on peat and coal for a large portion of its energy demands. It is likely that this reliance will continue for several more decades. Coal is generally brown to black and consists of an aggregate of mainly organic constituents, called macerals, which are associated with minor to moderate amounts of mineral matter (Karayiğit & Köksoy, 1994; Taylor et al., 1998). Differences between types of coal depend on depositional and biological variations, as well as on different degrees of coalification. With increasing pressure and temperature during burial, peat changes with time into brown coal, then into bituminous coal, and finally into anthracite (Fig. 1). During this coalification, hydrocarbons are generated and expelled biogenically and thermogenically.

Upper Carboniferous coal deposits, which underlie most of the Netherlands, are the main source rock for natural gas accumulations, including the giant Groningen field with its initial recoverable reserves of ca. 2700 $\times 10^9$ m^3 of gas (mainly methane; De Jager & Geluk, this volume). Part of the generated methane is not expelled but is retained within the coal itself. Depending on the process that releases the gas from the coal, it is referred to as coal-mine methane or coalbed methane.

Peat

Geological setting

In general, peat is formed under anoxic conditions in swamps, marshes or fens where organic matter from decaying vegetation is produced faster than it is decomposed by biochemical processes. Peat formed in large parts of the Netherlands during the Holocene (De Gans, this volume). In the coastal lowlands of the country (northern Groningen, Friesland, Noord-Holland, Zuid-Holland) it is generally referred to as low peat, and in the higher areas in the northeast (Drenthe, eastern Groningen, eastern Overijssel) and south (Peel: Noord-Brabant, Limburg) as high peat (Fig. 2). Applied in relation to present-day groundwater level, however, this subdivision has no genetic meaning (Engelen, 1987). In the west, large-scale peat formation occurred mainly between 2500 BC and 0 AD in a coastal to lower delta-plain setting. This peat is generally of the nutrient-rich bog-type. On the higher Pleistocene substrate, extensive nutrient-poor moorlands dominated (Van Montfrans et al., 1988). The raised dome structures of these high peats may have been more than 5 m above the surrounding environment. Nowadays, small patches of domed peat have been preserved only in a few scattered places (Engelen, 1987).

Historical development

During the early Middle Ages, peat was produced from the higher grounds of the northeastern Netherlands, especially in Drenthe and southeast Groningen, as well as in northwestern Germany and northern Flanders. This 'turf' (Dutch for a brick-sized block of dried peat) was used predominantly for heating purposes. From the 13th to the 16th century, production gradually shifted from northern Flanders and western Brabant to the northern Netherlands

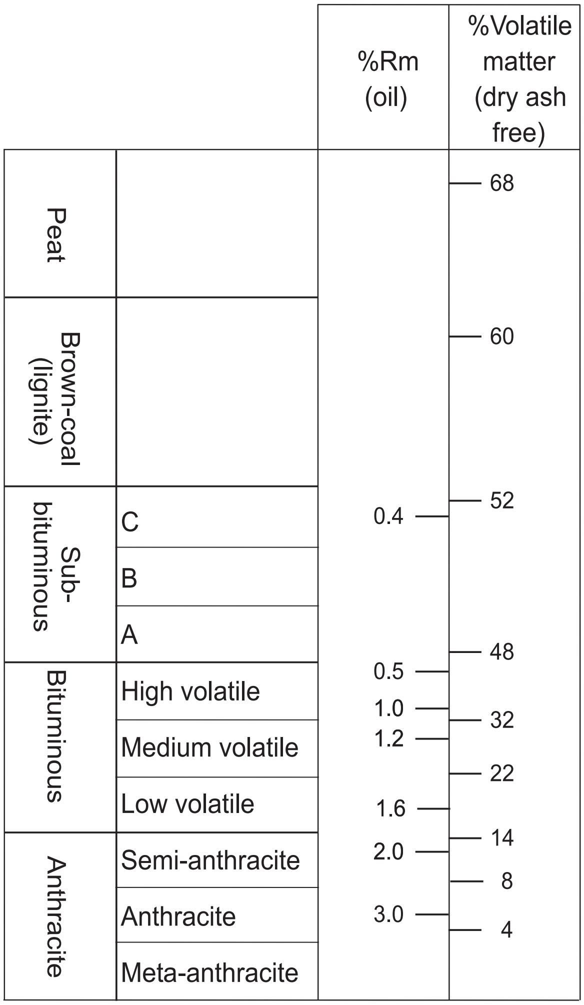

Fig. 1. Overview of ASTM (American Society for Testing and Materials) coal-rank classification (after Stach et al., 1982, and Taylor et al., 1998). The stage of coalification is indicated by the rank of the coal, most commonly expressed in percentage vitrinite reflectance (%Rm) or in percentage of volatile matter (daf = dry, ash-free)). Coalification is the progressive change in composition and structure of peat deposits, and any other organic carbon components within sediments, during burial, as a result of various interrelated physical, chemical and biological processes (Levine, 1993). The nature of the processes changes significantly with the different stages of coalification (Stach et al., 1982; Taylor et al., 1998). With advancing coalification, biochemical processes become less active. Anaerobic microbial processes have some influence until the formation of brown coal. During further burial, only physical factors (pressure, temperature) play a role (Van Krevelen, 1993). Brown coal has a higher density, lower porosity and lower moisture content than peat. During further coalification, brown coal changes into bituminous coal, and finally into anthracite, as a result of loss of water and thermally controlled chemical alteration of the organic matter. These processes result in volumetric shrinkage due to physical compaction.

(Fig. 3). Between 1550 and 1950, about 1000 km^2 of high peat and 420 km^2 of low peat in Friesland, Groningen, Drenthe and Overijssel were excavated for the production of turf. This turf is estimated to represent 3.5×10^9 Gigajoules of energy (Gerding, 1995). The lowering of the ground level by these large peat-cutting activities led to increased surface subsidence as a result of drainage, compaction and oxidation of the remaining peat layers. This, in combination with the gradual but steady rise of the water table due to continuous sea-level rise during the Middle Ages, caused the bogs in the west of the country to be drowned (Van Montfrans et al., 1988). The combined effect of erosion through marine incursions and the human exploitation of peat formed many lakes in the western and northwestern parts of the country.

In the second half of the Middle Ages, farmers exploited extensive peat lands in the western Netherlands for fuel and agriculture. In the higher areas, peat production continued in combination with an increased exploitation of wood from the last remaining large forest areas (Joosten, 1989).

Between 1700 and 1900, most of the remaining high moors and raised bogs in the northeastern Netherlands were drained, loosened, and subsequently burned by farmers to create suitable lands for the cultivation of buckwheat (Zagwijn, 1986).

In the Netherlands and surrounding areas, major turf production gradually stopped between the two world wars. Based on annual production data it is estimated that 9.4×10^9 m^3 of turf was produced from 1200 to 1950 (production from the Peel area not included; Leenders, 1987). Nowadays, peat is used in horticulture and the activated-carbon industry. It is produced for the latter purpose since 1921 in the Amsterdamse Veld and Schoonebeeker Veld, both located in southeast Drenthe (Engelen, 1987).

Brown coal

Geological setting

The brown-coal, or lignite, layers in the province of Limburg are part of the extensive brown- coal deposits of the Ville and Inden formations that were formed during the Miocene in the Lower Rhine Embayment north of the Eifel and Ardennes uplands (Fig. 4). Brown-coal deposits reach a maximum thickness of 108 m in this embayment in Germany (Van der Burgh et al., 1988). The Miocene Breda Formation in Zuid-Limburg consists of unconsolidated sand and envelops three to four intercalated brown-coal beds (Fig. 5; Engelen, 1987; Wong et al., this volume). The individual brown-coal beds may reach thicknesses exceeding 15 m (NITG, 1999). The coal beds continue to the west and northwest into Belgium and Noord-Brabant, where their thickness decreases markedly.

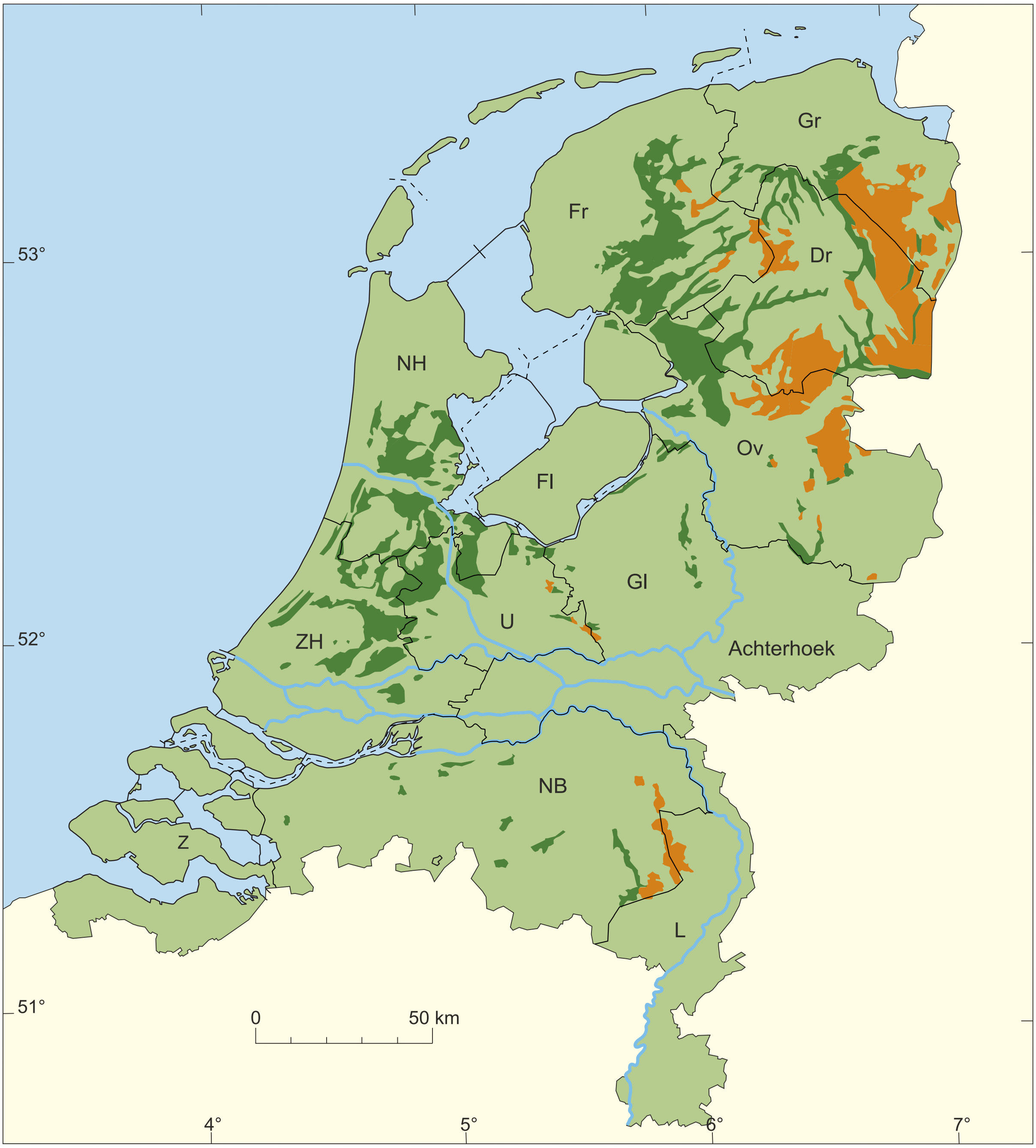

Fig. 2. Distribution of Holocene high (orange) and low (green) peat in the Netherlands. Compiled after Oosting (1937) and Geologische Dienst (1951). Most of the high peat shown has been excavated. The patchy distribution of low peat in the west results to a large extent from erosion. Provinces: Dr: Drenthe; Fl: Flevoland; Fr: Friesland; Gl: Gelderland; Gr: Groningen; L: Limburg; NB: Noord-Brabant; NH: Noord-Holland; Ov: Overijssel; U: Utrecht; Z: Zeeland; ZH: Zuid-Holland.

Historical development, resources

In Germany, in the area between Cologne, Bonn and Aachen, small-scale brown-coal mining started during the 16th century. In Zuid-Limburg, small-scale, though illegal, exploitation commenced between 1865 and 1868 near Eygelshoven (Engelen, 1989a). The first mining concession was granted in 1906 near Heerlerheide, but the exploitation was not successful (Engelen, 1987). It took until 1917, when fuel became scarce during the First World War, before new brown-coal extraction started in several villages near Heerlen. Dutch brown-coal production peaked at about 1.5 Mt/a around the end of the war (1918–1919).

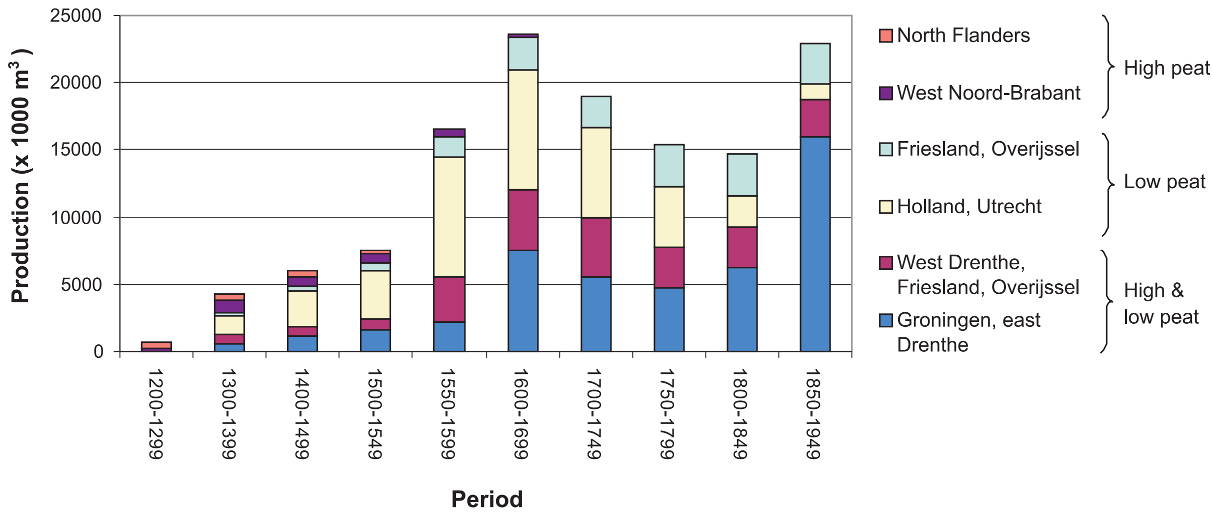

Fig. 3. Volumes of peat produced in the Netherlands and the north of Flanders, per period and per region. Please note the differences in the time periods. Figures indicate average yearly production and do not include the Peel area (east Noord-Brabant and northern Limburg; after data from Leenders, 1987).

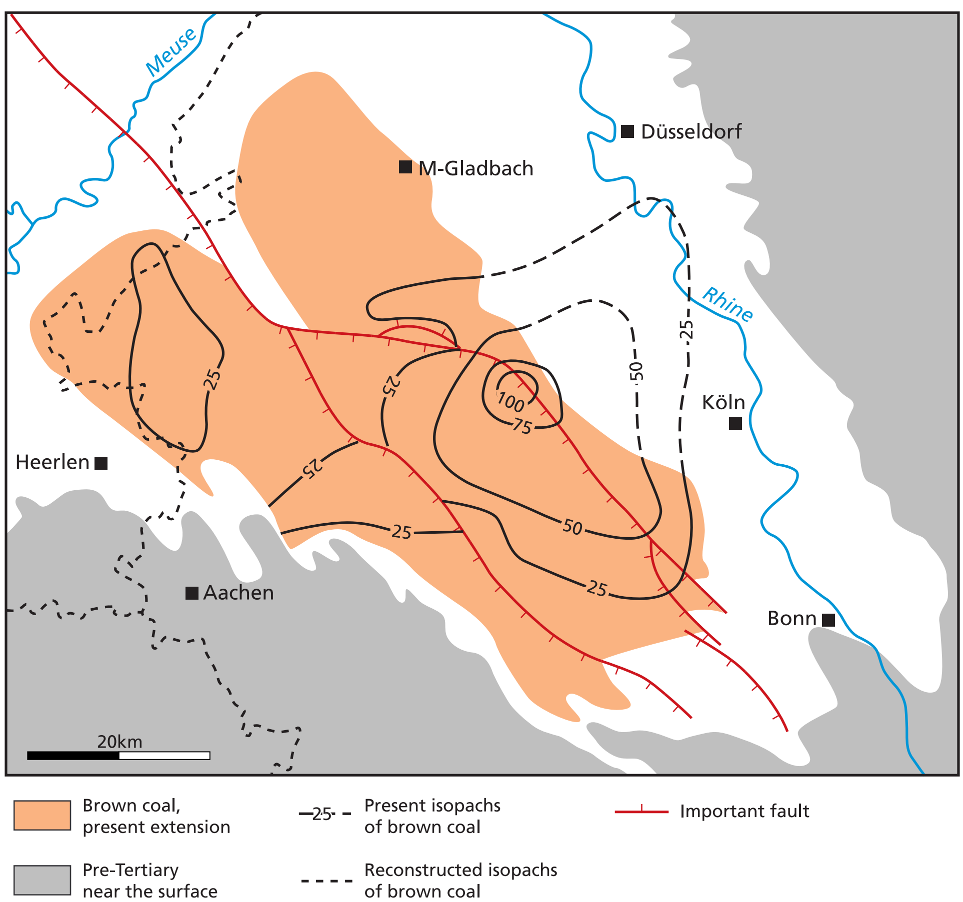

Fig. 4. Map showing the occurrence and thickness (metres) of Miocene brown-coal deposits in the Lower Rhine Embayment (after Zagwijn & Hager, 1987).

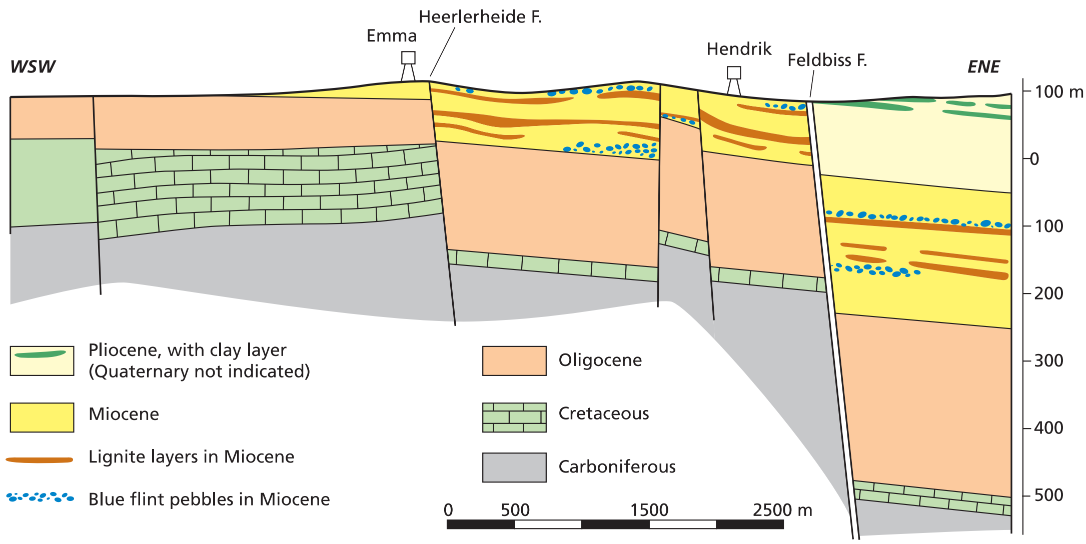

Fig. 5. Cross section through Tertiary strata near Heerlen in northeastern Zuid-Limburg, showing the depth of the brown-coal-bearing Miocene silver sands (after De Jong & Van der Waals, 1971). For location of the Emma and Hendrik mine shafts see Fig. 6.

In 1921, production had decreased to less than 0.13 Mt. In the following years most quarries ceased operations. Renewed production commenced during the Second World War and lasted until 1962. It reached over 0.2 Mt/a during these last years (Engelen, 1989a). The brown-coal resources in the Netherlands are calculated to exceed 1700 Mt (Van der Burgh et al., 1988). However, despite their relatively shallow depth, any future mining is unlikely because of environmental concerns and population density.

The brown coal of the Lower Rhine Embayment in Germany continues to be mined intensively, providing cheap fuel for the large power plants in the area between Cologne and Aachen, not far from the border with the Netherlands. It is mined in very large open pits to a depth of 500 m. The annual brown-coal production is over 100 Mt, and the original recoverable reserves in place were estimated to be about 50 000 Mt (Van Montfrans et al., 1988; D. Juch, personal communication). Nowadays, the recoverable reserves are estimated roughly at about 10 000 Mt (D. Juch, personal communication). During mining, millions of cubic metres of groundwater are pumped off and discharged. As a consequence, groundwater flow directions can be reversed, and formerly wet areas on the order of 10 to 100 km wide are drying (Stuurman, 2000; NITG, 2001; De Vries, this volume).

Bituminous coal and anthracite

Geological setting

The main coal-bearing formations in the Netherlands were deposited during the Late Carboniferous (Westphalian) in the Northwest European Coal Basin. During a phase of deformation and uplift at the end of the Carboniferous, about 2000 m of Carboniferous strata were eroded across large parts of the country, leading to the current configuration of the Carboniferous subcrop below the Base Permian Unconformity (Pagnier et al., 1987; Van Buggenum & Den Hartog Jager, this volume). At present, the coal-bearing deposits are found at shallow depth in the southernmost Netherlands and deepen northward to more than 3 km.

The Namurian Epen Formation consists mainly of lacustrine, marine, and deltaic claystone with some sandstone intervals that locally are overlain by coal beds. The main coal-bearing deposits belong to the Baarlo, Ruurlo and Maurits formations of Westphalian A to C age (± 318–309 Ma). These formations have a total thickness up to 3000 m, were deposited in a cyclic river-dominated environment, and have coal contents up to about 2% (Van Buggenum & Den Hartog Jager, this volume). The younger fluvial sandstone and claystone of the Westphalian C and D locally contain thin coal seams (Van Adrichem Boogaert & Kouwe, 1993–1997). The rank of the coal seams is mainly bituminous; anthracite occurs locally. Based on Markov analysis, parasequence thicknesses in the Westphalian deposits were quantified. Cycle thicknesses for the Westphalian A and B are around 10 m (Pagnier et al., 1987). However, not all cycles contain coal. The actual spacing between the seams is shown in Table 1. Based on five coal-exploration wells drilled in the 1980s in the Achterhoek (eastern Gelderland) and in Zuid-Limburg, the mean thickness of Westphalian coal seams of 50 cm or more in thickness is 1 m (RGD, 1986). According to Van Bergen et al. (2000) the median thickness

Table 1. Spacing and thickness of Westphalian coal seams (≥ 50 cm thickness) in the areas shown in Fig. 8 (Zuid-Limburg: depth < 1500 m; Peel: < 1500 m; Achterhoek (including south Overijssel) < 2000 m; Zeeland–Noord-Brabant: 1500–2000 m). Density of data varied per area. For a limited number of coal-exploration wells, the coal-seam thickness was known from cores; otherwise geophysical well logs (sonic, density) were used (from Van Bergen et al., 2000).

Area	Seam spacing					Seam thickness				
	Number of evaluated wells	Number of evaluated seams	Seam spacing (≥ 50 cm thick)			Number of evaluated wells	Number of evaluated seams	Seam thickness (≥ 50 cm thick)		
			minimum (m)	maximum (m)	median (m)			minimum (cm)	maximum (cm)	median (cm)
Zuid-Limburg	33	119	0.7	226.9	23.1	33	146	50	205	80
Peel	16	136	0.2	253.8	42.2	20	171	50	190	80
Achterhoek	25	110	1.7	473.4	35.1	25	133	50	295	88
Zeeland–Noord-Brabant	2	20	6.0	495.3	75.5	2	21	52	297	86

of such seams in the areas mentioned is 88 and 80 cm, respectively. Individual coal-layer thicknesses reach up to 3 m in the greater Achterhoek area (Table 1), or, according to some authors, even up to 3.5 m (RGD, 1986; Pagnier et al., 1987).

Historical development, recent resource evaluations

ZUID-LIMBURG

Good overviews of the history of coal mining in the Dutch province of Limburg until 1974 are given by Visser & Zonneveld (1987), Stuffken (1987) and NITG (1999). Much of the following text is based on these publications.

The lower part of the Westphalian coal measures is at or near the surface in the valley of the Worm river north of the German city of Aachen, where it has been mined from Roman times onwards in simple open pits. Underground mining in galleries started early in the 14th century. In the 16th century, when exploitation began to require more technical and financial resources, an industry developed that was based on the early capitalist system (Raedts, 1971; Engelen, 1989b). At the beginning of the 18th century, the abbey of Kloosterrade (Rolduc) began developing coal mines. By the middle of that century, mining was conducted from small shafts. During the French occupation of 1794–1815, mining came to a near standstill as a result of seizure and incompetent management. During the French period, all natural resources became the property of the State under the Napoleonic law of 1810. Under this law, concessions for exploitation were granted by the State. Exploitation in the state-owned mines was unsuccessful after 1815. The Domaniale concession was therefore leased to a private enterprise in 1845. Coal mining became truly established at the end of the 19th century when the advent of a railway line improved access to eastern Zuid-Limburg. During this period, the large Oranje-Nassau concession was awarded, and the Oranje-Nassau I mine became operational in 1899.

From 1902 onwards the 'State Collieries' (Staatsmijnen) Wilhelmina, Emma, Hendrik and Maurits were established in the substantial areas not yet covered by earlier concessions. The new concessions were repeatedly extended in the first half of the 20th century. Smaller enterprises opened mines along the German border.

Coal mining in Limburg took mainly place between the Benzenrade Fault and the First NE Main Fault (Fig. 6). As a result of the north to northeast dip of bedding and because of faulting, the depths at which exploitation took place varied considerably. The bottom of a shaft of the Domaniale mine was only 22 m below surface level, whereas a shaft of the State Colliery Hendrik extended down to 1008 m. A total of 248 coal seams occur with a cumulative coal thickness of over 73 m (IGCP 166, 1980). Approximately 65 of these seams are exploitable, with thicknesses ranging roughly from 50 to 150 cm (Kuyl, 1980).

After the global economic crisis in the late twenties and early thirties, the Dutch coal-mining industry experienced a boom, interrupted by a wartime low, that lasted until after World War II (Fig. 7). Twelve surface facilities were in operation, with a combined annual production of about 12.6 Mt (Engelen, 1989b). In the early 1960s, when production was peaking, the mines provided work for more than 28 000 workers underground and nearly 23 000 at the surface (Westen, 1971). Interest in coal outside Zuid-Limburg diminished rapidly after the discovery of the gi-

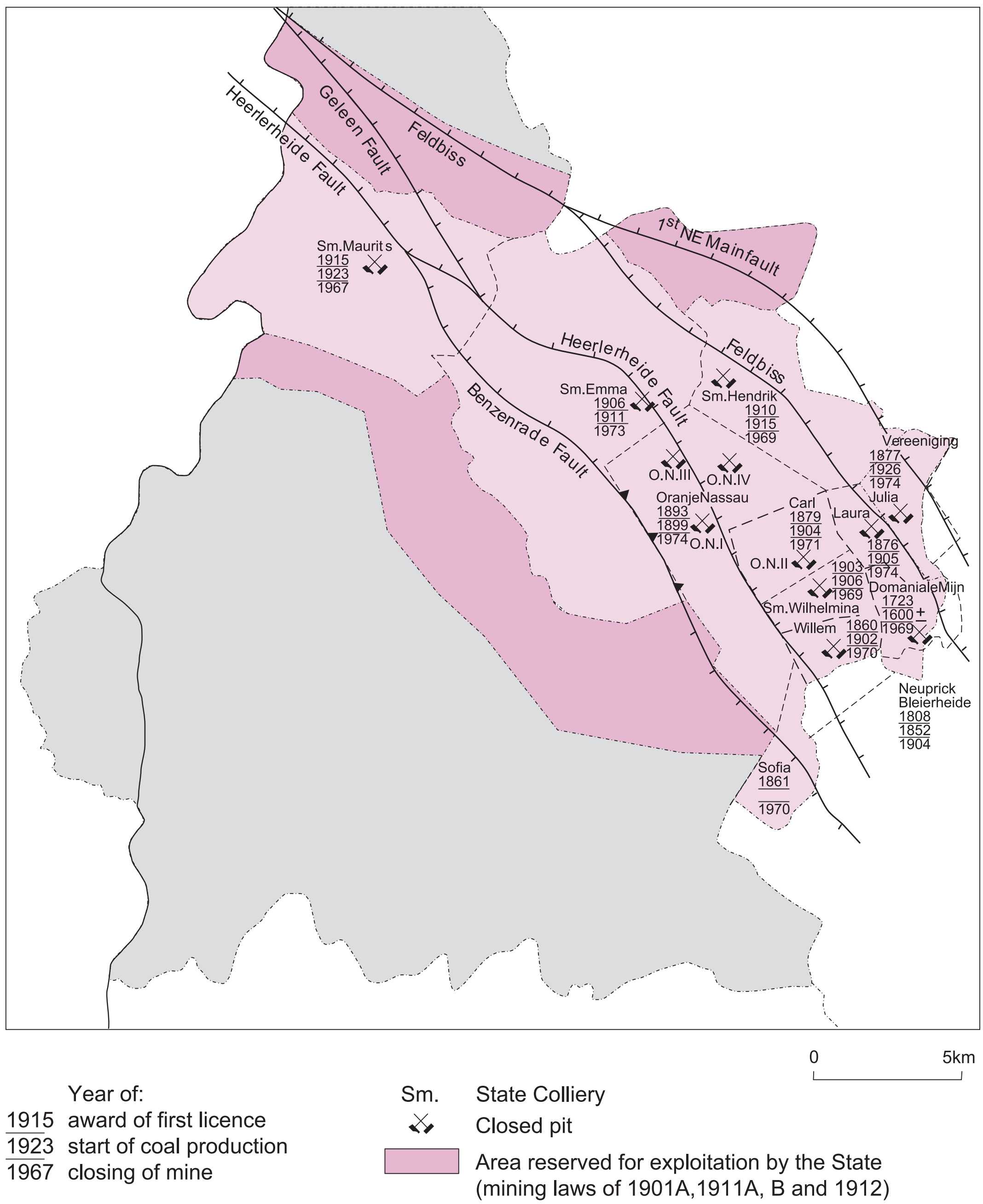

Fig. 6. Overview of the Zuid-Limburg mining district, showing major faults, locations of mines and coal concessions in 1960, with details on the history of exploitation (from NITG, 1999; after Raedts, 1971).

ant Groningen gas field in 1959 (Van Tongeren, 1987). This discovery, together with low world-energy prices and rising exploitation costs, led to the collapse of coal mining in the Netherlands (Raedts, 1971; Van Tongeren, 1987). The period from 1967 to 1974 saw the rapid closure of the mines.

In the 20th century, approximately 570 Mt of coal were mined in the Netherlands (Westen, 1971; Fig. 7). Yet, enormous coal resources remain in Limburg and elsewhere. In the Peel and Achterhoek areas, several exploration wells for coal were drilled but no mining has occurred (Peelcommissie, 1963; RGD, 1986). During the 1980s, renewed interest in coal, induced by the second oil crisis and also triggered by new techniques such as underground coal gasification, led to an inventory of coal deposits down to a depth of 1500 m (RGD, 1986; Pagnier et al., 1987; Van Tongeren, 1987). The results of this inventory indicate that an approximately triangular fault block of ca. 20 km² between the

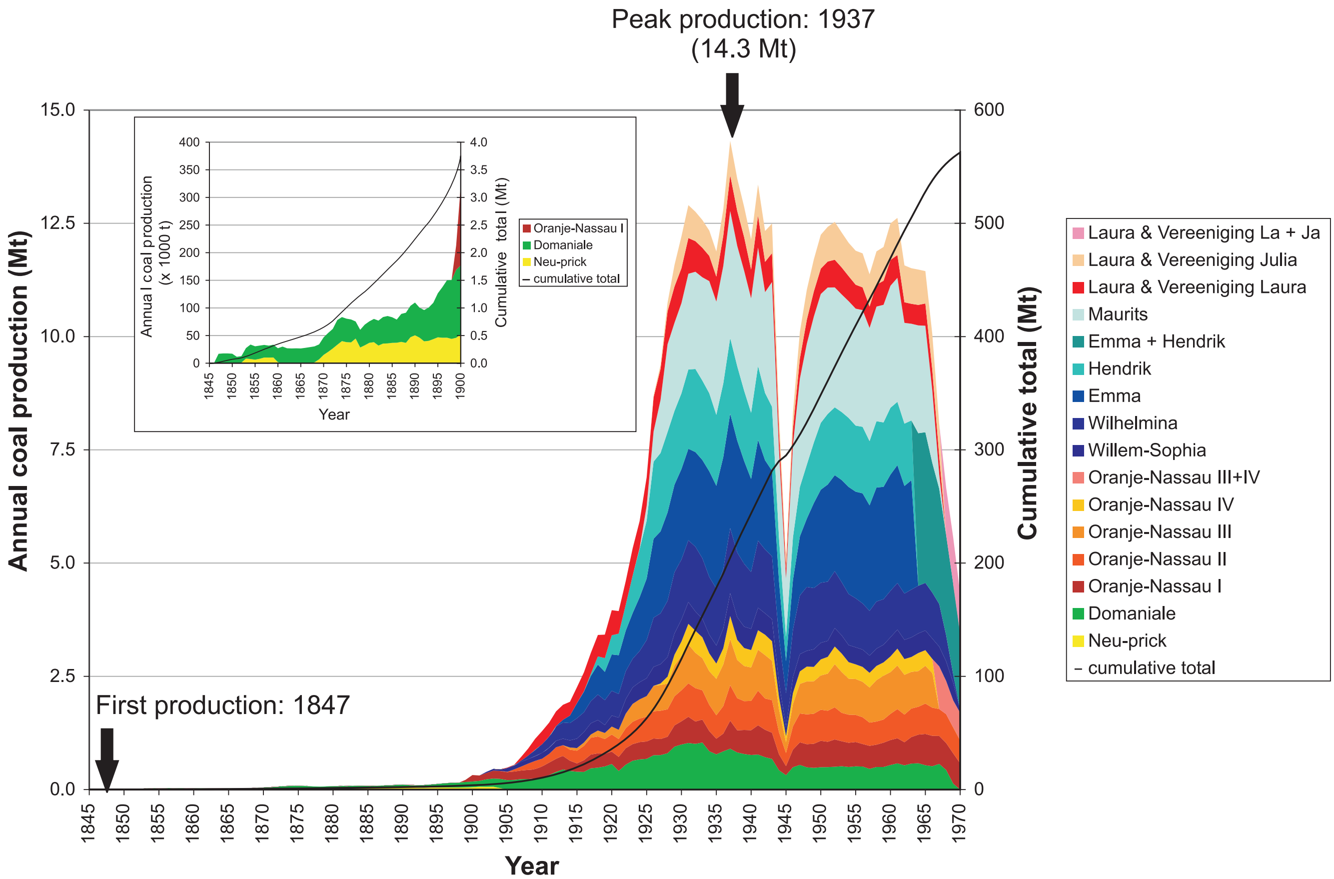

Fig. 7. Net annual coal production from 1847 to 1970 from mines in Zuid-Limburg (after data of Westen, 1971). Data up to 1929 without coal sludge; since 1929 including coal sludge. Extrapolation of the declining production curves indicates a cumulative production of ca. 8.5 Mt in the period from 1971 until the closure of the last mine in 1974.

Heerlerheide Fault and the Feldbiss Fault, north of the former mining area, contains a coal resource of 591 Mt (RGD, 1986). Underground coal gasification was considered as early as 1946 (Visser, 1987), but no application resulted, although elsewhere, especially in the USA, demonstration projects have been conducted.

A new inventory of the coal resources in the Netherlands down to a depth of 1500 m in Zuid-Limburg, the Peel and the greater Achterhoek area (including south Overijssel) was finalised in 2000 (Fig. 8; Van Bergen et al., 2000). Coal below this depth is unlikely to become of economic significance in the near to mid-term future. Within all three areas, faults with throws of at least 50 m were mapped using seismic data. In Zuid-Limburg, the mined fault blocks were excluded from the inventory. The coal resources per block were calculated based on the dimensions of the block, the net cumulative coal thickness, and the coal density (Van Bergen et al., 2000). In the Ruhr Basin in Germany, coal seams are fairly constant in thickness over large areas, and variations in individual coal seams are compensated by variation in other coal seams (Drozdzewski, 1992; Süss et al., 2002; D. Juch, personal communication). A similar pattern probably exists in the Netherlands, but could not be confirmed in view of the lack of coal-seam correlations between scarce wells over large distances in the investigated areas. It was therefore decided to follow a probabilistic approach. The expected resources per block are presented as a cumulative probability curve to take into account the uncertainties in the thickness and density of coal (Table 2, Fig. 9). The calculated quantities represent resources and thus include unrecoverable coal. The coal resources in the unmined part of Zuid-Limburg are estimated to be about 1580 Mt (P50 value; Table 2, Fig. 10a).

PEEL

In view of the coal resources immediately across the border in Germany, the State Service for the Exploration of Mineral Resources (ROD) carried out a drilling campaign between 1903 and 1916 to investigate the occurrence of exploitable quantities of coal in the Peel area. Based on the results from 13 deep wells and many shallow boreholes, the ROD calculated a reserve of about 2500 Mt of coal above 1500 m below sea level in seams of exploitable

Table 2. Estimated coal resources, potentially producible CH_4 volumes and potentially storable CO_2 quantities per area evaluated (Van Bergen et al., 2000). The CH_4 volumes of the Zuid-Limburg, Peel and greater Achterhoek areas are corrected for their burial histories. No correction was made for Zeeland–Noord-Brabant since this area was considered to be close to its deepest burial.

Area	Total area (km^2)	Depth interval (m)	Coal resources (1000 Mt)			CH_4 (10^9 m^3)			CO_2 storage (Mt)		
			P90	P50	P10	P90	P50	P10	P90	P50	P10
Zuid-Limburg	67	< 1500	1.0	1.6	2.3	1.7	3.5	6.8	8.1	17.4	34.7
Peel	383	< 1500	1.6	3.4	5.2	2.2	5.5	11.5	14.3	36.9	77.6
Achterhoek	4078	< 1500	1.4	9.1	30.4	3.3	23.4	88.4	16.4	118.4	481.8
	4078	1500–2000	1.4	9.1	30.2	7.3	50.2	190.1	39.8	260.0	1029.0
Zeeland–Noord-Brabant	2458	1500–2000	16.0	36.1	58.7	46.9	117.9	243.9	221.2	589.9	1229.0
Total		< 1500	4.0	14.1	37.9	7.2	32.4	106.7	38.8	172.8	594.1
		1500–2000	17.4	45.1	88.9	54.2	168.1	434.0	261.0	849.9	2258.1

Fig. 8. Areas and sub-areas, outlined in red, of which the coal resources, coalbed-methane resources and CO_2 storage potential were inventoried by Van Bergen et al. (2000). The main areas with Westphalian coal-bearing deposits shallower than 1500 m are Zuid-Limburg, the greater Achterhoek area (including southern Overijssel) and the Peel. In the Zeeland–Noord-Brabant area (including southern Zuid-Holland) and the major part of Overijssel, Westphalian coal seams occur at depths between 1500 and 2000 m, and deeper.

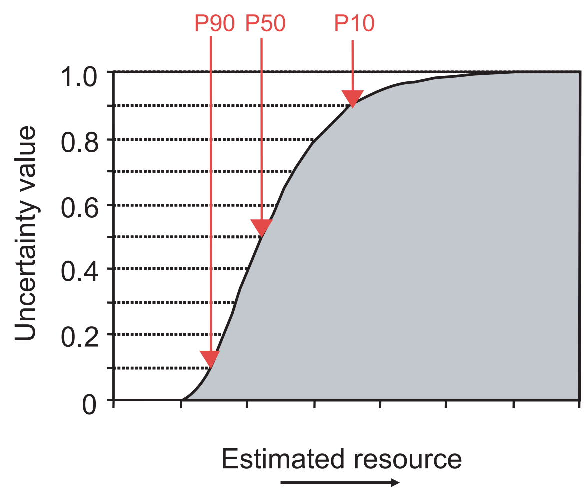

Fig. 9. Schematic cumulative probability curve, representing the result of a Monte Carlo analysis for resource estimation. This analysis is applied to take account of the uncertainties in the parameters, and allows predicting the probabilistic distribution of the expected resources. The estimations at probabilities P90, P50 and P10 carry uncertainties of 10, 50 and 90% respectively (cf. Table 2).

thickness (Van Waterschoot van der Gracht, 1918). During the Second World War, a gravimetric survey was carried out in the Peel and in adjacent areas (De Sitter, 1949; Van Weelden, 1957). Halfway through the 20th century, the State Collieries drilled 12 wells (NITG, 2001). In 1952, exploration intensified after the establishment of an advisory committee, the Peelcommissie. It included the shooting of 368 km of seismic data and the drilling of eight additional wells. The committee calculated that a geological reserve of approximately 2900 Mt of coal in seams over 50 cm thick occurs in the evaluated area shallower than 1500 m, and observed that the net cumulative coal thickness in this area is less than in Zuid-Limburg (Peelcommissie, 1963). Meanwhile, the construction of two shafts for the new Beatrix State Colliery in the eastern Peel area commenced in 1952; it was subsequently suspended in 1962 as a result of a change in energy policy (Peelcommissie, 1963; NITG, 2001). In 1984, further investigations were carried out by the Geological Survey (RGD, 1986; Pagnier et al., 1987). Seven seismic lines with a total length of 115 km were shot, covering about 220 km^2. The results of the re-

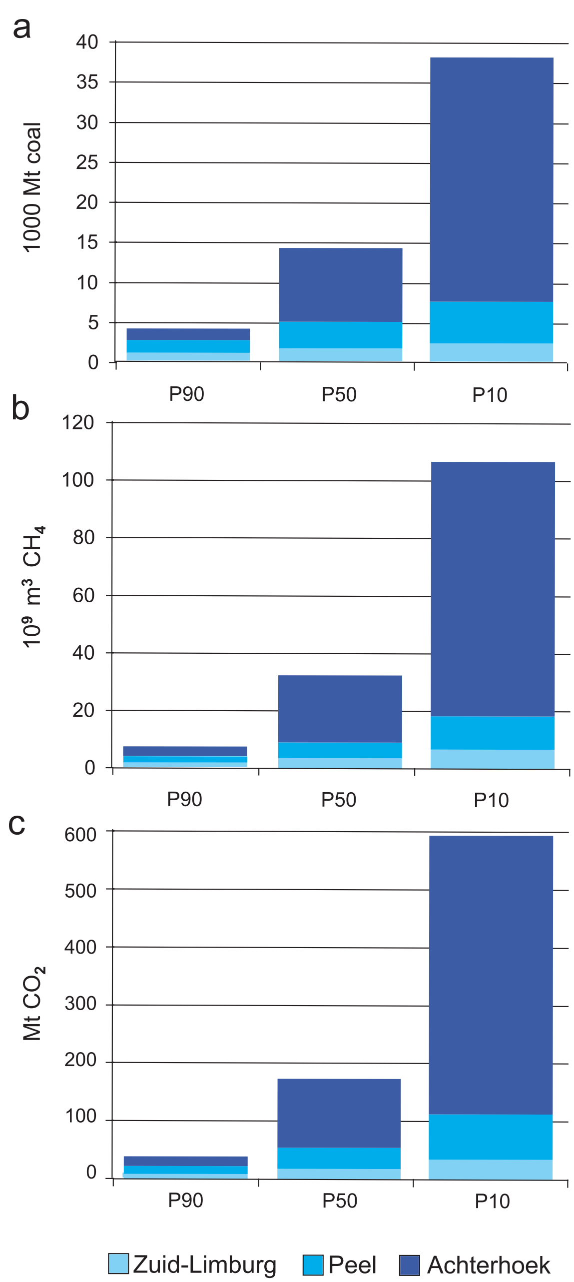

Fig. 10. Calculated quantities of (a) coal resources, (b) potentially producible coalbed methane and (c) storage potential for CO_2 in the Netherlands in Westphalian coals above 1500 m depth (see data in Table 2). The 'Achterhoek' area includes southern Overijssel (Fig. 8).

search confirmed the conclusions of the Peelcommissie. They also showed that the Carboniferous to the north of the Peel area is represented by less productive coal seams of the Namurian and Westphalian A (NITG, 2001).

For the recent evaluation of the Peel area, twelve fault blocks were distinguished, some of which are located in Germany, but fall under a Dutch concession according to an agreement reached in 1959 (Peelcommissie, 1963). The resources of the area (P50: 3360 Mt; Fig. 10a, Table 2; Van Bergen et al., 2000) are slightly higher than those reported earlier by the Peelcommissie (1963).

Achterhoek (and southern Overrijssel)

A concession for coal and rock-salt exploitation in the Achterhoek was granted in 1930, after the presence of Upper Carboniferous coal measures had been demonstrated near Winterswijk by the State Service for the Exploration of Mineral Resources. The area of this 'Gelria' concession is greater than 100 km². At that time, the reserves of bituminous coal in this area between 900 and 1500 m depth were estimated at about 360 Mt. However, mining never took place because of high costs and risk involved, not to mention an economic recession (Visser, 1987).

In the 1980s, further investigations were carried out by the Geological Survey (RGD, 1986; Pagnier et al., 1987). Within the framework of that study, seismic lines with a total length of 79 km were shot and 375 km of previously shot lines were evaluated. The resources of three coal seams from Top Carboniferous to 1500 m, in a triangular area of 157 km² between the coal-exploration wells Joppe-1, Hengevelde-1 and Ruurlo-1, were estimated at 693 Mt (RGD, 1986).

The total coal resources shallower than 1500 m of the greater Achterhoek area, including south Overijssel, are estimated to be about 9100 Mt (P50; Fig. 10a) according to the recent inventory by Van Bergen et al. (2000). Only blocks bounded by faults with offsets larger than 100 m were considered due to relatively limited data.

Comparison of resources between the areas

The recent inventory shows that net coal thickness is highest in Zuid-Limburg but that total resources in this relatively small area are smaller than in the Peel and greater Achterhoek areas (Van Bergen et al., 2000). The uncertainty in the calculated resources is lowest in Zuid-Limburg because of its coal-exploration and mining history. The large, greater Achterhoek area potentially contains the largest coal resources (Table 2, Fig. 10a). However, only a fraction of these resources can be considered proven (P90), and the major part of the resources occurs between depths of 1000 and 1500 m. The resources of the area between 1500 and 2000 m are in the range of 9170 Mt (P50).

Although poorly known, coal resources between 1500 and 2000 m depth are also expected to exist in northern Zeeland, southern Zuid-Holland and western Noord-Brabant (Fig. 8). In this structurally stable area on the north flank of the Brabant Massif, only a few seismic lines are available, and the area was evaluated as a single block. Total coal resources for this area are estimated at 36 080 Mt (P50; Table 2). This figure, however, results from the extrapolation of data from a few wells over relatively long distances, and therefore carries a large uncertainty.

Coalbed methane

Introduction

Coalbed methane, or CBM, is the natural gas that is retained in coal in the subsurface. It is known as mine gas or coal-mine methane in the mining industry, where it is dangerous due to its explosive nature. From the 1970s onwards, interest has grown in coalbed methane as a fuel. Due to a concerted effort by the United States government and private organisations to demonstrate commercial production, the coalbed-methane industry in the United States grew from a little-known, high-cost operation to a competitive, main-stream natural-gas producer (Saulsberry et al., 1996; Ayers, 2002).

Many European countries possess important coal resources, which are not suitable for conventional mining because of economic or technical reasons. The Carboniferous deposits in western Europe show many similarities with those of the Black Warrior Basin in Alabama, also a Variscan foredeep basin, which is, after the Cretaceous San Juan Basin in New Mexico and Colorado, the most productive coalbed-methane area in the United States (Pashin, 1998). During the 1990s, interest in coalbed methane was growing in Europe. Fails (1996) compared the coalbed-methane potential of five Variscan-foredeep coal basins in Germany and Great Britain with the Black Warrior Basin. Compared with the latter basin, the European foredeep basins tend to contain much thicker coal measures with more numerous coal beds and greater net coal thickness. Fails (1996) also compared coal rank, methane content, potential depth of production and post-coalification histories. The unmined Eastern Ruhr Basin in Germany, one of the basins evaluated, appears to be attractive for the production of coalbed methane with high exploration risk but potentially high economic return. However, production in Europe appears to be sub-economic so-far.

No correlation was found between coal thickness and production performance in the Black Warrior Basin (Pashin & Hinkle, 1997). In fact, the most productive wells in that basin are in relatively thin coal (Pashin & Hinkle, 1997; Pashin et al., 2004). The factors that control coal-gas producibility are hydrodynamics, depositional setting and coal distribution, tectonic history and structural setting, coal rank and gas generation, permeability, and gas content (Scott, 2002).

Hydrodynamic conditions strongly affect coalbed-methane producibility and include both the movement of meteoric water basinward, important for biogenic gas production, as well as the migration of fluids from deeper parts in the basin. Coal beds are the source and reservoir for methane, and their widespread distribution within a basin is critical to establishing a significant coalbed-methane resource. Sufficient gas should be generated through time from the coal, either through thermogenic or biogenic processes. Thermogenic gas generation is strongly linked to burial depth, and therefore to the tectonic history of the basin. The structural style of the basin is important for the continuity of the coal seams. In tectonically active areas, coal seams were often buried deeper than they are at present. This paleo-burial depth affects, as a result of irreversible compaction during burial, present-day porosity and permeability. The reservoir properties of a coal bed are far more complex than those of a conventional gas reservoir such as sandstone. Permeability in coal beds is determined by their fracture (cleat) system, which is in turn largely controlled by the tectonic regime (Scott, 2002). Permeability strongly depends on present-day in-situ stress directions. Compressional stress directions orthogonal to face cleats will lower permeability, whereas those parallel to face cleats may enhance permeability. Additionally, precipitation of authigenic material may reduce permeability (Scott, 2002; Pitman et al., 2003). The most productive wells have permeabilities ranging between 0.5 and 100 milliDarcies. Because permeability decreases with depth, coalbed-methane production may be limited to depths less than 1500 to 1800 m (McKee et al., 1988; McCants et al., 2001; Scott, 2002). A permeability that is too high may have a negative impact because it results in high water production.

Most of the methane (> 90%) in a coal is adsorbed on the coal's microstructure. The gas content in the water-filled pores and cleats of the coal is minimal. The gas-sorption capacity of coal is generally interpreted to depend on pressure, temperature and coal characteristics (Bustin & Clarkson, 1998). The relationship between pressure and sorption capacity appears to follow a pressure-dependent Langmuir isotherm. This isotherm displays a parabolic increase of the gas-sorption capacity with increasing pressure, until a maximum is reached above 20 MPa (Fig. 11). This behaviour reflects monolayer adsorption on a surface, where the maximum represents a completely saturated surface along which the monolayer approaches liquid density. The methane content calculated with the Langmuir isotherm is consequently the maximum that the coal seams can contain at the in-situ reservoir pressure. With increasing temperature, the gas-sorption capacity shows a decrease that is reported to be either linear (Kim, 1977; Levy et al., 1997; Bustin & Clarkson, 1998) or exponential (Yang & Saunders, 1985). Consequently, the opposed effects of pressure and temperature result in a maximum in gas-sorption capacity at a certain depth (Fig. 11). Initially, down to 2000 m, pressure will have the largest effect; below that depth, temperature is likely to be-

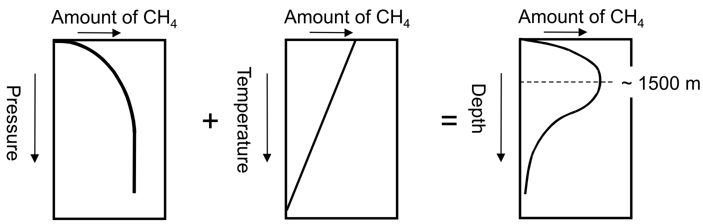

Fig. 11. Effects of increasing pressure, temperature and depth on the amount of methane adsorbed on coal, assuming a normal geothermal gradient and hydrostatic pressure comparable to those in the Carboniferous of the Ruhr Basin in Germany (Freudenberg et al., 1996). This figure shows a lineair decrease (after Kim, 1977; Levy et al., 1997; Bustin & Clarkson, 1998); an exponential decrease was reported by Yang & Saunders (1985).

come dominant. The maximum gas-sorption capacity occurs in general around a depth of 1500 m.

Coal characteristics that affect sorption capacity are maceral composition, rank, ash content and moisture content. The influence of maceral composition on sorption capacity is complex, and might be related to the pore structure of the macerals (Ettinger et al. 1966; Clarkson & Bustin, 1996; Gamson et al., 1996; Crosdale et al., 1998; Lamberson & Bustin, 1993; Levine et al. 1993; Beamish & Gamson, 1993; Crosdale & Beamish, 1993). Poor or no correlation between sorption capacity and maceral composition is also reported (Faiz et al., 1992; Laxminarayana & Crosdale, 1999; Bustin & Clarkson, 1998). Coal rank is generally considered to be the main parameter affecting the methane-adsorption capacity of coal (Ryan, 1992). In general, the sorption capacity increases with coal rank (Eddy et al., 1982; Yee et al., 1993, Laxminarayana & Crosdale, 1999), although U-shaped relationships between sorption capacity and rank are also reported (Moffat & Weale, 1955; Patching & Mikhal, 1986). These relationships may be related to pore-structure development with rank (Levine et al., 1993). However, on a global scale there are statistically no significant linear or non-linear correlations between rank and adsorption capacity (Bustin & Clarkson, 1998). The ash content of coal is generally considered to act as a simple non-sorbing diluent, thereby reducing the storage capacity (Bustin & Clarkson, 1998; Laxminarayana & Crosdale, 1999). The moisture content of coal has, in general, a negative effect on sorption capacity (Joubert et al., 1974; Seewald & Klein, 1986; Yalcin & Durucan, 1991; Levine et al., 1993; Levy et al., 1997; Bustin & Clarkson, 1998; Mavor et al., 1990; Lama & Bodziony, 1996). Moisture in coal competes for adsorption sites on the coal surface and may block access of gas to the microporosity (Yee et al., 1993; Bustin & Clarkson, 1998; Laxminarayana & Crosdale, 1999).

Given the above, the actual in-situ gas content is often difficult to assess on the basis of isotherms alone. For example, the coals are often found to be undersaturated with methane, i.e. their gas content is lower at a certain pressure and temperature than can be expected on the basis of the isotherm. Undersaturation of coal can be the result of uplift, possibly resulting in degassing of coal beds, and of cooling during geological history with no additional thermogenic gas being generated (Scott et al., 1994; Fig. 12). McCants et al. (2001) relate the undersaturation observed in coal seams in the Upper Silesian Basin in Poland to degassing. Degassing probably also occurred in the Ruhr Basin in Germany. Undersaturation due to cooling can be the result of decreasing basinal heat flow or to a temperature decrease during uplift. In volumetric calculations, the burial history of the area has to be considered to be able to estimate the paleo-hydrostatic pressure after the latest uplift event, and to correct for a possible degassing (Van Bergen et al., 2000).

Investigations in the Netherlands

Following the successes in coalbed-methane production in the United States, the well Peer was drilled in 1992 in the Campine Basin in northeastern Belgium, near Zuid-Limburg. This is the only well close to the Netherlands that was drilled and tested for coalbed-methane production. Only a limited amount of CH_4 was produced because the inflow of water into the well was too high to reduce the pressure sufficiently for substantial gas desorption. These unfavourable hydrological conditions were due to the proximity of a major open fault (the 'Donderslagbreuk'). Wenselaers et al. (1996) concluded that, despite the low productivity observed, perspectives for coalbed-methane production exist in the Campine Basin, with more than enough coal and gas present. Nevertheless, no further wells were drilled until present.

The coal basins in the Netherlands resemble the Eastern Ruhr Basin in that the coal-bearing deposits are of the same age and the basins are located in a similar position with respect to the Variscan front. Fails (1996) established that the depositional setting and coal distribution of the Eastern Ruhr Basin are favourable for coalbed-methane production. The depositional setting and coal distribution in the Netherlands are also considered to be suitable for such production. The tectonic history and structural setting are in general more complex in the Netherlands than in coalbed-methane producing basins in the United States. The Dutch coal-bearing sequences are rather heavily faulted and locally also folded, especially near the Variscan front (e.g. in Zuid-Limburg). Still, tectonic fault blocks of substantial size with more or less continuous coal seams can be identified. Most coal seams have been buried deeper than their present depth. The Westphalian coal in the Netherlands is bituminous or anthracitic, indicating that large volumes of thermogenic gas have been generated during coalification. Contributions from secondary biogenic gas generation are so far

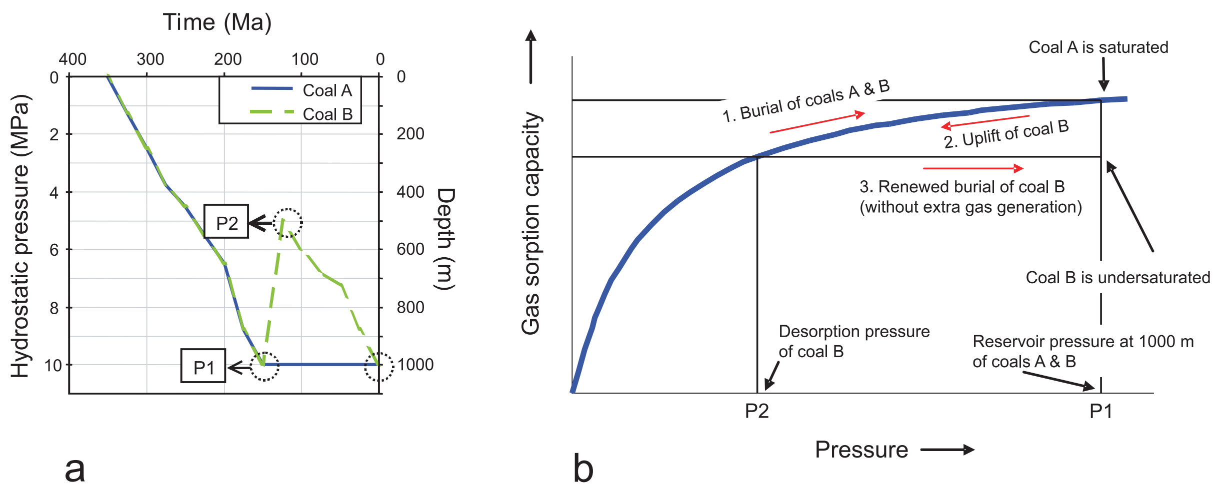

Fig. 12. The effect of burial history on the amount of adsorbed methane. (a) Coals A and B were buried to depths of 1000 m. While coal A remained at that depth, coal B experienced a phase of uplift, after which it was reburied to 1000 m. (b) During the uplift, coal B released some of its adsorbed methane. Assuming no renewed gas charge, coal B will not adsorb more methane during reburial, and hence becomes undersaturated (adapted from McElhiney et al., 1993).

not reported and seem unlikely due to the depth of the coal. In coal at around 1000 m, a permeability of 0.1 to 1 (maximally about 5) milliDarcies is to be expected in the Netherlands. The permeability in the aforementioned Peer well was less than 1 milliDarcy in Westphalian A and B seams at depths between 850 and 1250 m, i.e. about 450 m above their maximum paleo-burial depth (Wenselaers et al., 1996). In view of the lack of coal-mining activities in the Netherlands since 1974 and of coal exploration since 1985, the gas content can only be estimated by indirect methods. Stuffken (1957) investigated the methane content of coals in Zuid-Limburg on the basis of data from mine-ventilation air and found a U-shaped relation between coal rank and gas content, with a maximum at about 25 wt% volatile matter. However, it is difficult to convert these data from ventilation air into gas content per in-situ coal volume. Coppens (1967) showed that the sorption capacity of coal samples from the Belgian part of the Campine Basin varies, depending on rank, between 5 and 9 m^3/t. A small set of isotherms, recently obtained on core samples from wells in the Achterhoek area, shows the sorption capacities of the investigated dry coals to be about 15 m^3/t, while those of moist coal were lower. No rank effect could be discerned (Krooss et al., 2002). Probably, this lack of correlation between rank and gas content is due to the small number of samples.

Although production is not yet economically feasible in the Netherlands, the resources of coalbed methane were inventoried in several studies. Examples from the Black Warrior Basin and the Rocky Mountains coal basins show that saturated coal exists primarily in recharge-driven fresh-water plumes, where late-stage bacterial methanogenesis facilitates high gas contents (Scott, 2002; Pitman et al., 2003). Identifying recharge areas and subsurface flow paths is important for delineating prospective coalbed-methane plays (J.C. Pashin, personal communication). The general high salinity of formation water in the Dutch Carboniferous deposits, however, does not indicate recharging with meteoric water. Unfortunately, the data to perform a proper evaluation of the hydrological conditions in these deposits are currently lacking and should be derived from future well information. The inventory studies are therefore based mainly on isotherm data and on relations between rank and gas content.

An estimation by Wolf et al. (1997) indicated very high resources: 770 × 10^9 m^3 down to a depth of 2000 m and 1400 × 10^9 m^3 in deeper seams. However, the negative impact of increased temperatures below 2000 m was not taken into consideration, and the resources are therefore probably overestimated. The amount of gas that can potentially be produced, the technically recoverable reserves, depends on completion and recovery factors. The completion factor is the percentage of the in-place coal resources that will contribute to the gas production in the area to be developed. This strongly depends on the thickness and spacing of the coal seams. The completion factor will be high in thick, closely spaced seams, and low in thin, distantly spaced seams. The recovery factor is the percentage of gas that can be produced from the contributing coals. In coalbed-methane production, as applied in the USA, this depends strongly on the pressure drop that can be achieved in the coal seams by the pumping of large volumes of water; normally, 20 to 80% of the gas originally in place can be recovered. GAPS (1994) estimated the technically recoverable quantity of such gas in the Netherlands to range from 5 to 27 × 10^9 m^3 (NITG, 1999). The recent inventory by Van Bergen et al. (2000) focussed on the resources within Westphalian coal seams shallower than 1500 m, and calculated total recoverable resources of 32 × 10⁹ m³ for the three evaluated areas (P50 value; Fig. 10b). Van Bergen et al. (2000) used isotherms in their evaluation, which probably introduces, given the considerations above, an error into the estimates that is considered to be acceptable in the absence of additional data. The Zuid-Limburg area has potentially the highest producible methane content per square kilometre due to its high net cumulative coal thickness. The potential for total producible methane is highest in the large, greater Achterhoek area (including southern Overijssel) and fairly high in the medium-sized Peel area (Van Bergen et al., 2000).

In addition to the inventory down to 1500 m, methane resources for the depth interval between 1500 and 2000 m are large due to potentially large amounts of coal and high pressure. Estimated resources in the greater Achterhoek and Zeeland–Noord-Brabant areas amount to 50 and 118 × 10⁹ m³ gas respectively (P50). However, the producibility of these deeper resources is technically and economically questionable at present because the permeability of deeper coal seams is expected to be too low and the costs of drilling too high (Van Bergen et al., 2000; Scott, 2002).

The feasibility of coalbed-methane production in the Netherlands still awaits the evaluation of all production-related, technical and geological matters, as well as agreement on the societal, environmental, and economic aspects (NITG, 1999). In order to make coalbed-methane more attractive, it may be interesting to consider the combined exploitation of coal seams and (tight) sandstones within the Westphalian sequences. This type of completion is applied in the Uinta Basin in the western United States.

CO_2 sequestration

Introduction

The Kyoto protocol of 1997, in which many countries, including all members of the European Union, committed themselves to reduce their CO_2 emissions, recently opened new possibilities for the usage of coal. The Netherlands committed itself to a 6% reduction with respect to the emission level of 1990 in the period 2008 to 2012, i.e. to a total emission of not more than 160 Mt/a. The predicted emission under current policy equals around 210 Mt/a in 2010. This implies a reduction of 50 Mt in 2010 to reach the Kyoto target (Fig. 13). This quantity may become lower as the result of measures undertaken in the coming years to decrease energy consumption and to increase the use of renewable energy, but will remain substantial. Therefore, techniques are required to reduce emissions from fossil fuel. One of the options considered is to store CO_2 in subsurface coal beds. Gas adsorption has proven stable through geological time, and the risk of future CO_2 release is considered low. Injected CO_2 adsorbs at a near-liquid density firmly on the coal and replaces adsorbed CH_4. This released methane can be produced while injecting CO_2, with reduced production time and improved recovery compared to standard coalbed-methane production. Early laboratory experiments indicated that for two sequestered CO_2 molecules one CH_4 molecule is released (e.g. Puri & Yee, 1990; Stevenson et al., 1991; Hall et al., 1994). Since these early studies, several authors have presented different exchange ratios that are influenced mostly by pressure and coal rank. At higher pressures, in particular at depths below 800 m where the CO_2 is likely to be supercritical, the exchange ratio of CH_4 for CO_2 could be up to 5 : 1 at 12 MPa (~1200 m at hydrostatic pressure; Hall et al., 1994). Pashin et al. (2004) showed that the exchange ratio ranges from 2 : 1 in medium-volatile bituminous coal to 5 : 1 in high-volatile A bituminous coal at a pressure of 2.4 MPa. These observations were partly confirmed under laboratory conditions for Westphalian coal samples, with CH_4 : CO_2 exchange ratios varying between 1.2 : 1 and 6.3 : 1 (Wolf et al., 1999; Krooss et al., 2002; Van Bergen et al., 2000). For low-rank coals the exchange ratio can even be higher, exceeding 20 at lignite rank (Burruss, 2003). However, Busch et al. (2006) found that under certain conditions the exchange ratio is even less than 1, indicating preferential CH_4 sorption.

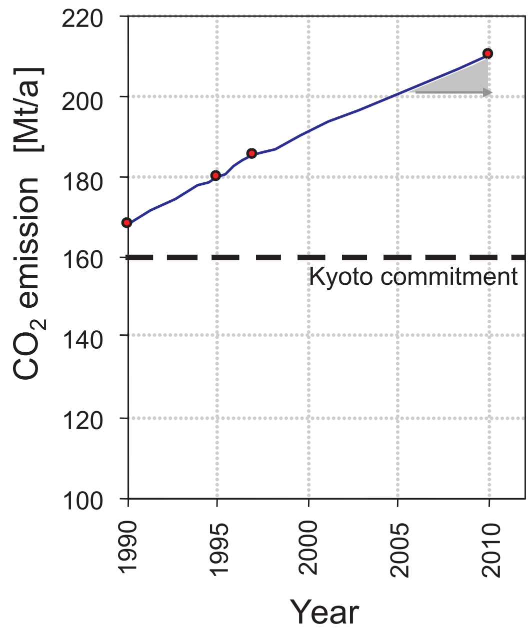

Fig. 13. Predicted total yearly CO_2 emission of the Netherlands until 2010, without reduction measures (Ministerie VROM, 1999). The shaded area indicates a reduction of 1 to 2 Mt/a that is required to stop the yearly growth of emissions. CO_2 sequestration in coal might play a role to achieve this reduction.

A high exchange ratio implies that, in areas rich in coal, power plants could be developed which emit less CO_2 per unit of energy produced than conventional plants, possibly even zero (Gunter et al., 1997), making this technology an exceptionally clean source of energy. Theoretically, CO_2 injection can increase coalbed-methane recovery up to 100% (Stevens et al., 1999). However, this will depend on the saturation of the coal; for the Black Warrior Basin the increase in recovery is estimated to be over 20% (Pashin et al., 2004). Other benefits of this technology are that coal seams can be exploited that are not minable, and that decentralised generation of energy in remote areas near coal basins can minimise the costs of energy transportation.

However, it must be emphasised that CO_2 storage in coal beds is not yet a well-established, mature technology. Although several theoretical studies illustrated the potential of the process, world-wide only a few demonstration sites have been installed. A micropilot field test was set up in Alberta, Canada (Gunter et al., 1997, 1998), a similar test was recently performed in China, two-well tests were realised in Japan and Poland (Van Bergen et al., 2006), and a multi-well pilot in the San Juan Basin in New Mexico, USA (Gunter et al., 1998; Erickson & Jensen, 2001; Schoeling & McGovern, 2000). The test in Poland showed that the injectivity of low-permeability coal could be increased substantially by hydraulically fracturing the coal and by using the proper injection equipment (Van Bergen et al., 2006).

Investigations in the Netherlands

The calculated amounts of CO_2 that can potentially be stored in coal seams in the Netherlands vary significantly in each area evaluated because of differences in hectarage and cumulative coal thickness; uncertainties, therefore, are large (Table 2, Fig. 10c). Estimated amounts of CO_2 that could be stored shallower than 1500 m range from 35 to almost 600 Mt (Van Bergen, 2000; Hamelinck et al., 2001). For comparison, the current total annual emission of CO_2 in the Netherlands is about 200 Mt. However, due to infrastructural, societal and environmental constraints it will be impossible to exploit all investigated areas. Nevertheless, the quantities of CO_2 that could be stored may be sufficient to play an important role in reducing the future growth of CO_2 emissions (ca. 1% or 2 Mt/a; Fig. 13).

An economic evaluation indicated that CO_2 sequestration in coal could be economically feasible in the Netherlands in the relative short term if the technology proves to be applicable (Hamelinck et al., 2001). However, incentives will be required to make this technology attractive for industry. Drilling costs will be a dominant cost factor. The best conditions for a test site are met in Zuid-Limburg, where gas content is relatively high and uncertainties relatively limited (Hamelinck et al., 2001). A feasibility study into the possibilities of developing a pilot project in Zuid-Limburg showed that at present enhanced coalbed-methane production in combination with CO_2 sequestration cannot compete with conventional natural-gas production, and that an incentive such as a CO_2 tax is required to make it economically feasible (Van Bergen et al., 2003).

Conclusions

The subsurface of the Netherlands contains enormous resources of fossil fuel in the forms of peat, brown coal, bituminous coal, and coalbed methane. Although it is currently not feasible to exploit these resources, coalbed methane could become attractive in the future when fossil fuel resources decline and technology advances.

Of particular interest is the potential of coal as a medium to store CO_2 while producing CH_4. In the years to come, the world will still depend on fossil fuel, despite all the effort put into the development of sustainable energy. Therefore, it is important to search for technology options that allow the continued use of fossil fuel with less CO_2 emission, and that can be applied on a large scale during the transition to sustainable energy systems. The option of CO_2 storage in coal with simultaneous methane production is economically not feasible under current conditions and still requires further technological research, but could become attractive in the near future in the context of international measures to reduce greenhouse effects.

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Salt

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ABSTRACT

In the Netherlands various salt deposits occur within the Permian and Triassic. Especially the salts in the Upper Permian Zechstein attain great thicknesses (up to 1000 m) in the north of the country and the adjacent offshore. This salt has been mobilized during geological history into a large number of salt diapirs and pillows and has strongly influenced the tectonic styles in the subsurface. The shallowest salt dome offshore is almost at the sea bed, onshore at a depth of around 100 m. Salt is produced by solution mining in the eastern and northern Netherlands: rock salt (halite, NaCl) from the Triassic Röt and the Permian Zechstein, and magnesium salt (bischofite, $MgCl_2 \cdot 6H_2O$) from the Zechstein. Total production amounts to 5.5 Mt/a of rock salt and 0.25 Mt/a of magnesium salt. Exploitation takes place at depths between 300 and 3000 m, the mining of Permian rock salt in the Barradeel concession being the deepest worldwide. The rock salt is mainly used for the manufacturing of chlorine; most of the $MgCl_2$ is used to produce magnesium oxide.

Keywords: Netherlands, rock salt, potassium-magnesium salts, halokinesis, solution mining, salt concessions

Introduction

Salt is part of a group of chemically deposited minerals which crystallize by evaporation of salt-rich lake or sea water. Their chemical compositions reflect the original composition of the brine. The largest known, and also economically most important salt occurrences on earth are of marine origin. The sequence of evaporites results from variations in the chemistry of the brine from which they were deposited. The word salt is used as a general name for evaporitic salts, like halite (NaCl), sylvite (KCl), carnallite ($KClMgCl_2 \cdot 6H_2O$), kainite ($KClMgSO_4 \cdot 2.75H_2O$), kieserite ($MgSO_4 \cdot H_2O$) and bischofite ($MgCl_2 \cdot 6H_2O$).

The Upper Permian Zechstein salt is the most widespread and thickest salt formation in the Netherlands. It contains rock salt, potassium-magnesium salts and thick sulphates (Geluk, this volume: a). Apart from the Zechstein, salt occurs in the Permian Upper Rotliegend Group, in the Triassic Röt, Muschelkalk and Keuper formations (Geluk, this volume: b), and locally in the Upper Jurassic Weiteveen Formation. The presence of thick Zechstein salt deposits in the subsurface plays an important role in the structural geology; areas with and without salt show different tectonic styles (De Jager, this volume). In petroleum geology, rock salt is important since it presents regional top seals for natural gas in Carboniferous, Permian and Triassic reservoirs (De Jager & Geluk, this volume). Furthermore, the heat conductivity of rock salt is higher than that of most other sediments. Especially in areas where extensive salt movement has taken place, the subsurface temperatures are highly differentiated. This affects the maturity of source-rock intervals and must be taken into account in basin modelling.

Salt is exploited in the Netherlands since 1919 by means of solution mining. Currently, in three concession areas rock salt is mined, and in one potassium and magnesium salts (Fig. 1; Table 1). In the adjacent part of Germany a conventional salt mine is operated in Borth, and cavern fields at Epe, Krummhörn, Nüttermoor and Xanten. These caverns are used for gas storage.

Fig. 1. Salt concessions in the Netherlands, and locations of caverns and the Borth salt mine in the adjacent part of Germany (after Sedlacek, 2002). In grey shading, concessions from which salt has been produced; in white, concessions from which no salt has been produced. Concession details are presented in Table 1.

Table 1. Salt production and salt concessions in the Netherlands (Fig. 1).

Concession	Owner	Salt formation	Depth	Annual production[1]
Adolf van Nassau[2]	Akzo Nobel Salt	Z2 Salt (diapir)	600-1600 m	2743 (2475) kt NaCl
Barradeel	Frisia	Z2 Salt (layered)	2500-3000 m	940 (1050) kt NaCl
Buurse	Akzo Nobel Salt	Röt Salt (layered)	200-300 m	[3]cumulative production 1440 kt NaCl
Gelria	Hope & Co.	Z1 Salt (layered)	600-1000 m	[4]
Twenthe-Rijn[5]	Akzo Nobel Salt	Röt Salt (layered)	300-500 m	1881 (1875) kt NaCl
Veendam	Nedmag	Z3 Salt (pillow)	1600-1800 m	225 (221) kt $MgCl_2$
Weerselo	Akzo Nobel Salt	Z1 Salt (complex)	400-1000 m	[4]

Depth refers to the actual or, in case no production has taken place, to the envisaged production depth.
[1] Production 2000 (1999).
[2] Including 'Uitbreiding Adolf van Nassau'.
[3] Salt production stopped in 1954.
[4] No salt production has taken place.
[5] Including 'Uitbreiding Twenthe-Rijn'.

Geology of salt

Salt is in many aspects unlike other sedimentary rocks. First, it can be deposited at high sedimentation rates (1-10 cm/a) and preserved in thick packages, up to 1000 times faster than siliciclastic sediments (Schreiber & Hsü, 1980). These rates increase with increasing solubility from gypsum ($CaSO_4 \cdot 2H_2O$) via rock salt to potassium-magnesium salts. Secondly, salt deforms easily in a visco-plastic, i.e. non-brittle, way. Thirdly, salt is usually a tight material, i.e. has a porosity of less than 1% and a permeability of 10^{-7} to 10^{-4} mD. Finally, near the surface the relative density of salt is higher than that of most other sediments, whereas below 500 m it is lower.

Salt deposition

Salt is deposited from concentrated brine, where the evaporation exceeds the inflow. It forms both in continental and marine settings. With evaporation, overall levels of concentration increase until the specific solubility of a mineral is reached. At this moment the mineral precipitates. The normal order of precipitation from marine brines starts with gypsum, followed by halite. The gypsum is already in an early stage dehydrated and converted into anhydrite. When most calcium and sodium have been precipitated, the saturation levels of potassium and magnesium salts are reached next, and sylvite and carnallite precipitate. If an excess of sulphate is still present, kieserite and kainite may precipitate at the same time. Particularly in the Zechstein, however, these minerals have been formed as a result of intensive chemical transformations (Borchert & Muir, 1964). The last mineral to precipitate is bischofite, which crystallizes from a brine of almost pure magnesium chloride. Fresh inflow of sea water may start a new sequence of salt precipitation. The precipitation of gypsum usually heralds such a new sequence; it separates for instance the different Zechstein formations from each other (Richter-Bernburg, 1955). In the Netherlands, five evaporite Zechstein formations (Z1-Z5) and a Zechstein Upper Claystone Formation can be distinguished (Geluk, this volume: a).

Salt tectonics

On a geological time scale, and when buried below 500 m, salt behaves in a visco-plastic way, whereas other sedimentary rocks show brittle deformation. Rock salt compacts already during early stages of burial to a tight mass with a constant density of 2168 kg/m³. Other sediments show an increase in density with depth owing to cementation and the reduction of pore volume as a function of overburden and pore pressures. Consequently, in near-surface positions, where sand and clay typically show densities of 1200 to 1400 kg/m³, halite is relatively heavy, whilst below 500 m it is lighter than surrounding rocks. This results in an unstable situation. Under the right conditions, the salt rises while solid rocks sink (Trusheim, 1963; Kockel, 1990; Remmelts, 1995). This process of flowing and rising salt is called halokinesis (Greek: `αλσ, salt; κινειν, movement). It will only start when the salt successions have thicknesses of at least several hundreds of metres, and when a fault with a significant throw affects the base of the salt. Without such a trigger no halokinesis will occur; this is for instance the case for the Zechstein in most of the province of Friesland, where a hardly disturbed 600-m-thick salt layer is present.

Salt will initially flow mainly in a lateral sense, forming what is known as salt pillows (Fig. 2). Above a pillow, the sedimentary cover is not yet pierced. In the next stage a pillow can evolve into a salt diapir. The growth of the pillow causes extension and subsequent faulting and weakening of the sedimentary cover above the evolving diapir (Fig. 3; Jenyon, 1986; Remmelts, 1996). In cases of strong uplift, erosion may further weaken this cover. These processes take millions of years.

The Zechstein salt in the Netherlands has formed many diapirs (Fig. 4). The boundary conditions for diapirism have been met on a wide scale in the geological past. There are several periods of structuration during which halokinesis has been triggered, such as a long period of extension from the Mid Triassic to the Early Cretaceous, and a phase of compression during the Late Cretaceous to Early Tertiary. Many Zechstein salt structures are related to large faults in the substrate; elongated salt walls, like those along the margins of the Dutch Central Graben, are related to long fault zones, whereas isolated, circular domes developed over intersecting faults (Fig. 4; Remmelts, 1995, 1996; De Jager, this volume).

The shapes and the internal structures of individual salt bodies are complex; reference is made to descriptions by De Boer (1971), Richter-Bernburg (1972, 1980), Kupfer (1976), Bornemann (1991) and Geluk (1995). Salt diapirs outcropping in the Iranian Dasht-e-Kavir (Jackson et al., 1990) and Zagros Mountains (Kent, 1979), and in the Spanish Pyrenees (Wagner et al., 1971) present analogues for buried salt structures in the Netherlands (Geluk, 1998, 2000).

A good understanding of the mechanical behaviour of the various components within a salt sequence is a prerequisite to understand their deformation. Some rocks, such as interbedded claystones, thick anhydrites (> 1 m) and carbonates are brittle in the subsurface. Potassium-magnesium salts on the contrary are mostly ductile during deformation. Mixed rock types behave differently than the separate components. For instance, a thin-bedded alternation of anhydrite and salt can be deformed in a ductile manner, while one thick layer of anhydrite embedded in salt deforms in a brittle way (Lotze, 1957). Usually the stiffer layers will be folded within the softer layers, a structure that can often be observed in conventional salt mines and also in salt cores from boreholes.

Salt structures and occurrence

Different internal tectonic styles can be recognized in salt structures. In literature a separation is commonly made between layers, pillows and domes or diapirs, which represent increasing states of salt deformation (Fig. 2; Trusheim, 1963).

In *salt layers* the top and bottom of the layer are more or less parallel to the interbedded elastic carbonate and anhydrite beds, reflecting the original deposition. Salt lay-

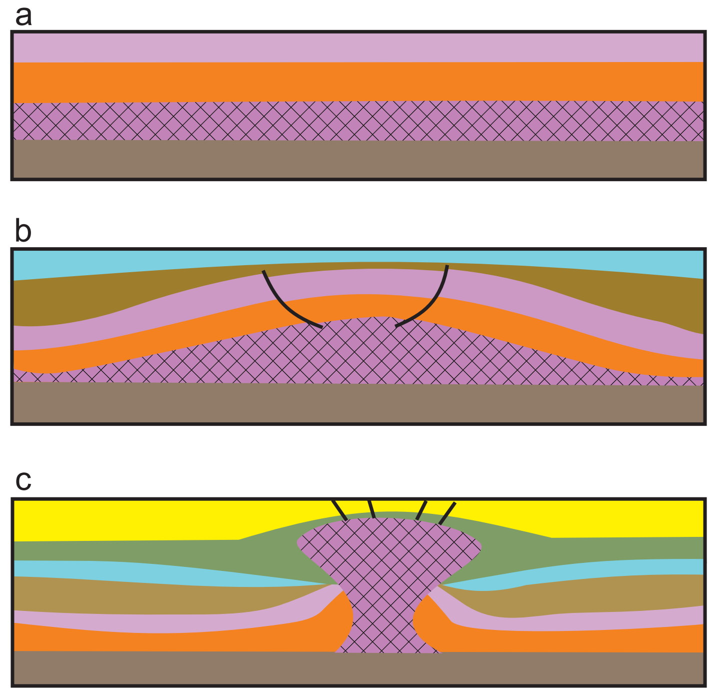

Fig. 2. Types of salt structures. (a) Layer, end Middle Triassic; (b) pillow, end Early Jurassic; (c) diapir, present-day situation. After Trusheim (1963).

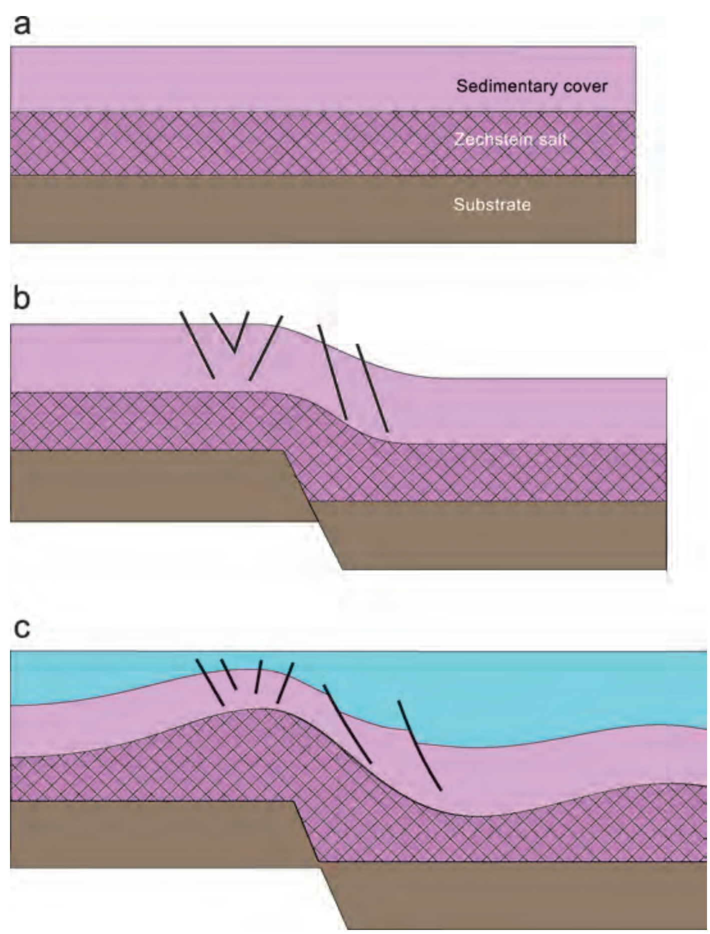

Fig. 3. Model for basement-fault-related salt movement in the southern North Sea. (a) Density inversion between the Zechstein salt and the overburden is created when the salt is buried below 500 m, but buoyancy forces are not large enough to overcome the strength of the overburden.
(b) Extensional faulting is accommodated by the salt layer, and the overburden is weakened due to stretching and faulting. (c) Differential loading in combination with tectonic stress enables the salt to flow. After Remmelts (1996).

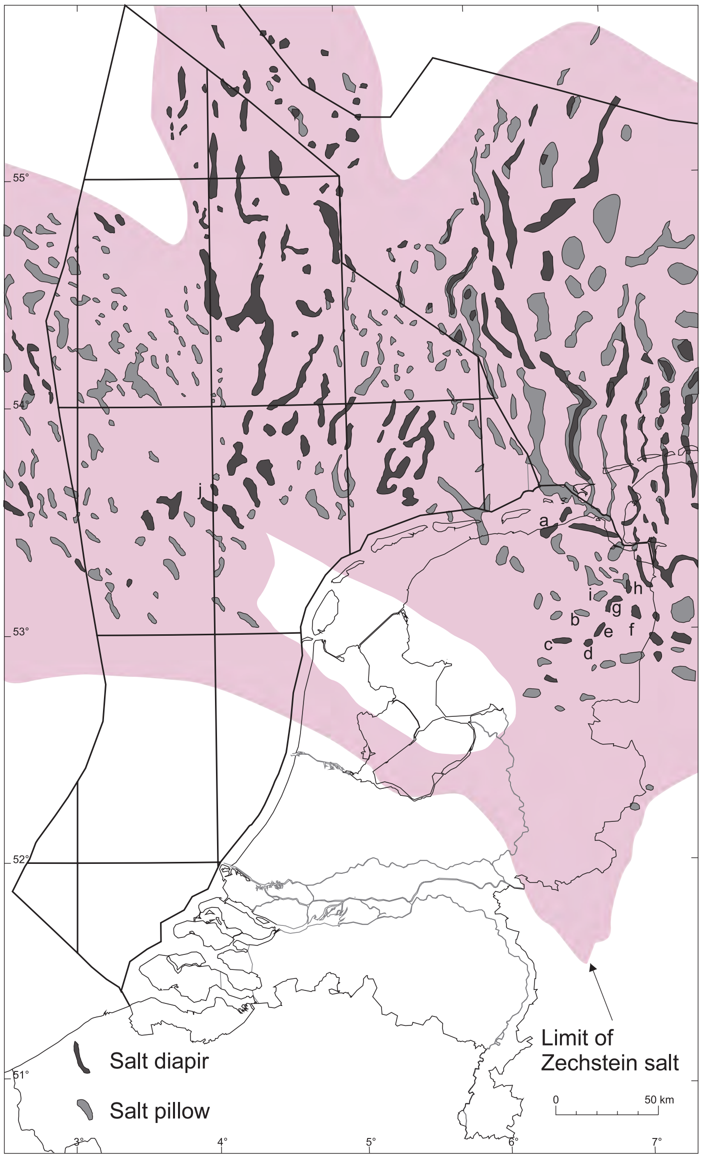

Fig. 4. Map of the Zechstein salt structures in the Netherlands (after Lokhorst, 1998). Elongated salt diapirs (e.g. in the northern offshore) are related to major sub-salt faults, whereas circular diapirs developed at intersections of fault trends. a: Pieterburen; b: Anloo; c: Hooghalen; d: Schoonlo; e: Gasselte-Drouwen; f: Onstwedde; g: Zuidwending; h: Winschoten; i: Veendam; j: K9.

ers may be faulted if the throw of sub-salt faults exceeds the thickness of the salt, e.g. in areas with relatively thin Zechstein salt. In basinal areas the thickness of the Z2 (Stassfurt) Salt (500-800 m) generally exceeds the throw of the fault, which then only affects the base. The internal structures of salt layers are mainly flow folds with horizontal axial planes (Lotze, 1957). They occur on the decimetre to decametre-scale. Potassium-magnesium salts, with interbedded layers of halite, usually show strong folding.

The salt occurrences in the Gelria, Twenthe-Rijn, Buurse and Barradeel concessions are examples of layered or bedded deposits (Fig. 1; Table 1). In the Gelria concession, up to 300 m of rock salt occur in the Z1 (Werra) Formation, at a depth between 600 and 1200 m. In the Twenthe-Rijn and Buurse concessions, 50 to 100 m of rock salt occur in the Triassic Röt Formation at depths of 200 to 500 m (Harsveldt, 1980, 1986). In the Barradeel concession, bedded rock salt occurs in the Z2 (Stassfurt) and Z3 (Leine) formations. Only the Z2 Salt is being exploited. It has a thickness of up to 650 m and occurs at depths of 2200 to 3000 m, which makes it the deepest salt exploited in the world.

Salt pillows are formed by a local thickening of the salt due to horizontal salt flow. In central areas of the Zechstein basin the cores of the pillows are composed of the Z2 (Stassfurt) Salt; in marginal areas the Z1 (Werra) Salt has been mobilized into pillows. Extensional faulting commonly affects the beds over the crest of the pillow. The thickening of salt in a pillow is usually accompanied by a series of stacked, large-scale flow folds in the salt, with horizontal axial planes as has been reconstructed for the

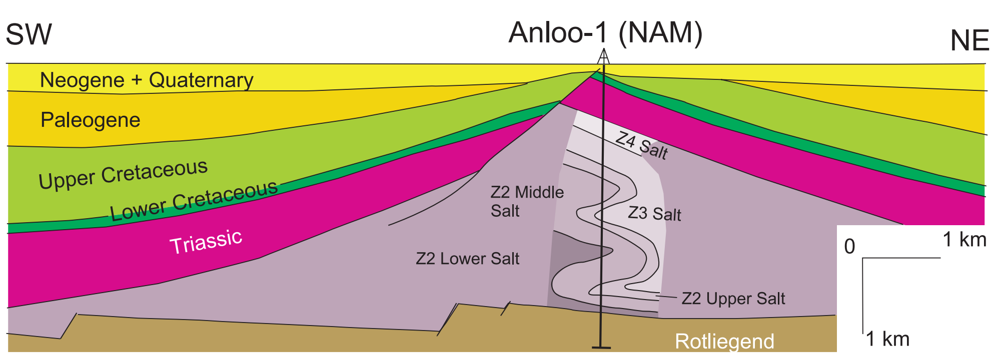

Fig. 5. Section of the Anloo salt pillow. Based upon the analysis of the wireline logs of well Anloo-1, several large-scale flow folds have been identified in this structure (after Geluk, 1995). See Fig. 4 for location.

Anloo salt pillow (Fig. 5). Interbedded carbonate and anhydrite layers within the pillow are folded and their axial planes are generally parallel to its roof. Superimposed upon such large-scale folds one commonly finds a series of smaller parasitic folds. In the late pillow stage, the direction of salt flow changes from horizontal to vertical (Lotze, 1957).

In the northern Netherlands, a series of Zechstein salt pillows occurs at relatively shallow depth (500-800 m). These pillows were brought close to the surface by deep erosion during the Late Jurassic to Early Cretaceous. In areas around the Dutch Central Graben, pillows of Permian (Rotliegend) as well as of Triassic (Röt, Muschelkalk) salts have been identified.

The Veendam concession covers two Zechstein salt pillows. Both contain thick layers of carnallite and bischofite in the Z3 Salt (Coelewij et al., 1978). It is the only known occurrence of primary bischofite in western Europe.

In *salt domes* or *diapirs* the movement of the salt changed from mainly horizontal to mainly vertical, and the salt pierced the cover beds. The vertical flow results in folds with vertical axial planes and horizontal fold axes as well as in curtain folds with vertical fold axes. As a result of the diapirism these latter folds usually have deformed the older generation of folds with horizontal fold axes. However, due to the extremely high strains of more than 300% in the diapir, the older deformation phase may be indistinguishable. Along the margins of diapirs, friction between the cover beds and the uprising salt mass results in high vorticities within this mass, with an additional complex fault pattern (Jackson et al., 1990). Sometimes the salt may even envelop blocks of the overlying succession. The salt in a diapir rarely moves at a uniform rate (Kupfer, 1976). Shear zones separate sectors moving at higher speed, known as spines, from those moving more slowly. The development of salt diapirs normally results in a complicated internal structure, with folds of many different scales and orientations (Richter-Bernburg, 1980). Compressional tectonics during the Late Cretaceous accelerated the rate of diapirism (Fig. 6), or strongly deformed some of the diapirs (Fig. 7). A generalized simplified model of Zechstein salt diapirs is that the central part of the dome contains mostly Z2 Salt and that the marginal parts are made up of younger salt units. This is shown to be true for structures as the Winschoten and Zuidwending diapirs (Harsveldt, 1980, 1986). In the Pieterburen diapir, however, the central part is made up of younger salts and the Z2 Salt occurs at the margin (Harsveldt, 1980; RGD, 1995); in Germany this occurs at Gorleben (Bornemann, 1991) and Hänigsen-Wathlingen (Schachl, 1987).

Several salt domes in the Netherlands pierced to close to the surface. The shallowest is the Zuidwending dome, where the top of the caprock is between 100 and 110 m below mean sea level (Fig. 6). Other shallow domes include Schoonlo (120 m), Pieterburen (190 m), Hooghalen (250 m), Onstwedde (250 m) and Gasselte-Drouwen

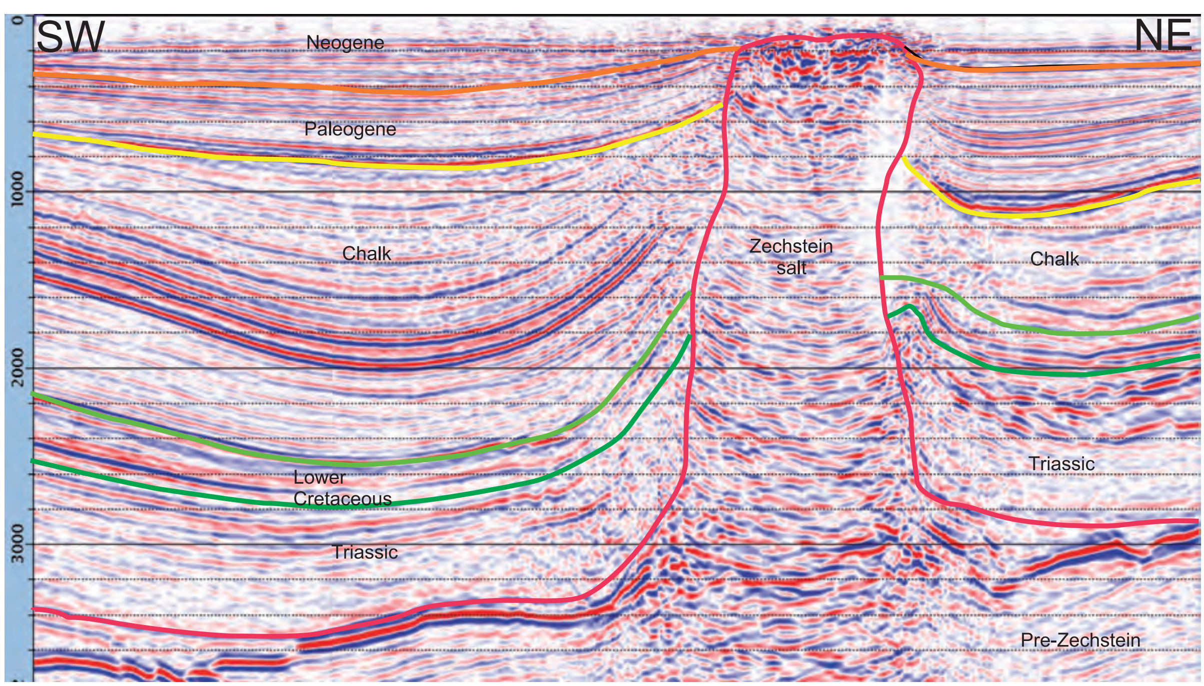

Fig. 6. Seismic section (depth in metres) of the Zuidwending salt dome in the Adolf van Nassau concession. This dome is the shallowest in the northern Netherlands, reaching to ca. 100 m below the surface. Thickness variations in the adjacent sediments result from periods of main salt movement. See Fig. 4 for location.

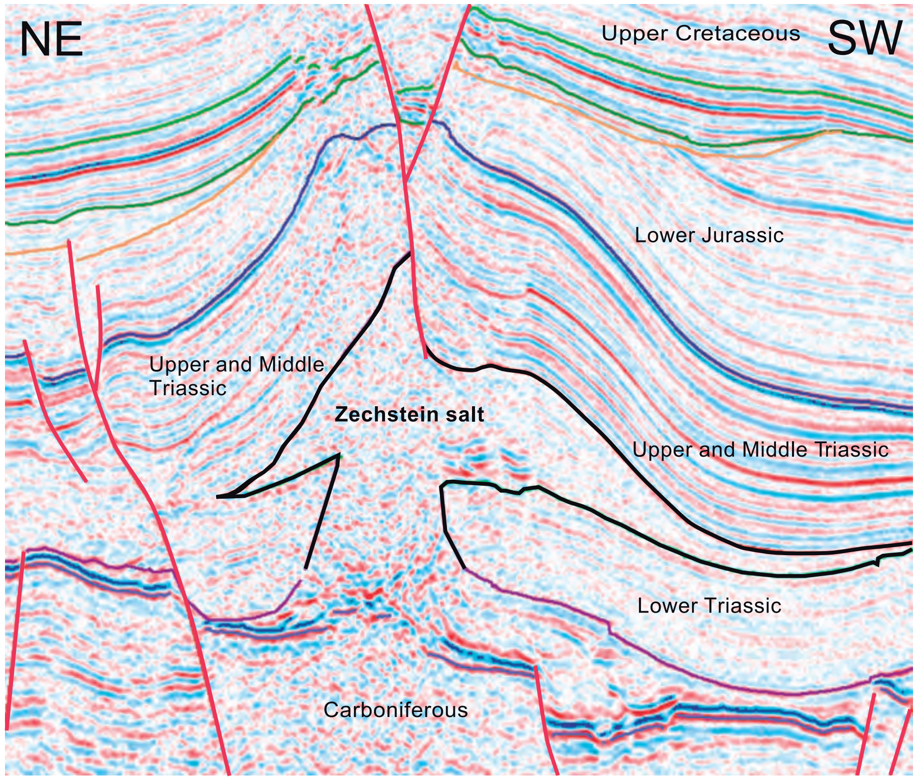

Fig. 7. Seismic section of a complex salt dome in the north of the Netherlands. This structure developed in response to Triassic faulting. Intrusion of Zechstein salt into the Triassic is related to Late Cretaceous compression (after NITG, 2000).

(350 m). In the western offshore block K9, a salt dome reaches almost to the sea bottom (Giesen & Mesdag, 1995).

Some of the Dutch salt domes have reached the surface at some time in the geological history, mainly during the Late Jurassic or Early Cretaceous, and became subject to erosion and subrosion, the near-surface dissolution of salt by ground or surface water. As a result a caprock formed, composed of the insoluble residue of the salt (Batsche & Klarr, 1980). This caprock may locally include thick anhydrite floaters. The thickness of the caprock shows considerable variation between salt domes, from 20 m in Winschoten to over 100 m in Pieterburen. Also within a single salt dome, this thickness may vary strongly in relation to the percentages of insoluables in the subcropping salt succession and their structural style (Balzer, 2000a, b, 2001).

In the Adolf van Nassau concession, two salt domes occur. The shallow Zuidwending dome has a typical banana-shape in map view, and is interpreted to be an amalgamation of two individual domes (Figs 6, 8). The top of the Winschoten dome is 400 m below the surface.

Salt exploration and production history

Rock salt was discovered in the eastern Netherlands during the 1880s in a drinking-water well (Wassmann & Brouwer, 1987). Later, during the drilling campaign of the Dienst der Rijksopsporing van Delfstoffen between 1903 and 1923, more deposits of rock salt and also of potassium-magnesium salts were discovered in this area (Van Waterschoot van der Gracht, 1909, 1918). In 1919, rock-salt production in the Buurse concession by solution mining of the Triassic Röt salt was started by the Koninklijke Nederlandse Zoutindustrie (KNZ, now Akzo Nobel Salt) following heavy taxes on German salt exports. In 1936 the production started in the Twenthe-Rijn concession close to the newly constructed Twentekanaal, which allowed better transport facilities. In the same period, the Gelria concession was granted for the exploitation of rock salt and coal. To protect the underlying Carboniferous coal reserves, the production of rock salt was restricted to conventional mining, which made it uneconomic. The production of rock salt in the Buurse concession was discontinued in 1953. During and after the Second World War several salt domes were discovered by drilling, others were suspected based upon gravity anomalies (Bentz, 1947; Mulder, 1950). More domes and pillows were detected by seismic surveying during the 1950s to 1980s. This resulted in the award of concessions for the production of rock salt (Adolf van Nassau and Weerselo) and potassium-magnesium salts (Veendam). The most recent concession, Barradeel, was granted in 1991 and enlarged in 2003 (Fig. 1, Table 1).

Salt production

The techniques applied in solution mining depend upon the geology of the salt and particularly on its internal structure. After a general introduction on solution mining, the mining of rock salt and potassium-magnesium salts will be discussed.

In the past, salt was produced from natural brine springs by evaporation. In the coastal areas of the Netherlands, it was produced by the burning of peat. Solution mining started by the pumping of natural salt-saturated water (Jeremic, 1994). In this process, which was never applied in the Netherlands, the produced water was replenished by fresh or unsaturated groundwater in an uncontrolled manner. As dissolution of salt occurred mostly at the top, the overlying rocks became unsupported. This resulted in subsidence at the surface. Large-scale collapse could even create a sinkhole. This type of dissolution may also occur naturally by groundwater flow, whereby salt is dissolved and flushed away.

In the Netherlands, controlled solution mining is applied, using several methods to regulate salt dissolution, raw brine production and cavern development. These methods involve considerations regarding cavern integrity and long-term stability, surface subsidence as well as post-abandonment monitoring.

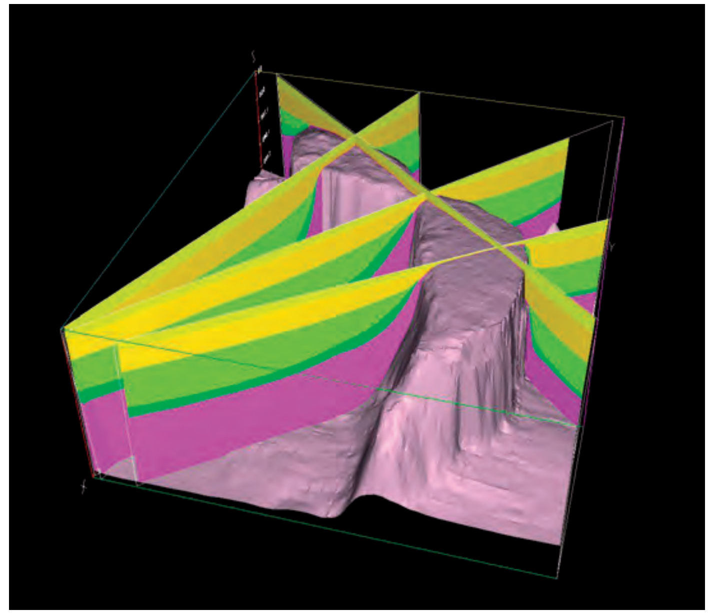

Fig. 8. 3-D model of the Zuidwending salt diapir. This structure is an amalgamation of two individual domes with a marked difference in strike. The height of the dome is ca. 3 km above the base of the Zechstein salt layer. The model covers an area of ca. 6 × 6 km. The view is towards the west. Yellow: Tertiary and Quaternary; light green: Upper Cretaceous Chalk; dark green: Lower Cretaceous; pink: Triassic. See Fig. 4 for location.

Rock-salt solution mining

Solution mining of rock salt may be designed as a single or multiple-well operation, depending upon the nature of the salt deposit and the shape of the cavern to be developed. Two modes of operation can be distinguished, i.e. the indirect and the direct circulation mode. The indirect circulation mode is also called 'top injection method', because the fresh water is injected in the upper part of the cavern. The water is flushed downward to the cavern bottom, increasing in salt concentration by dissolution of solid salt on its way. The brine is produced near the base of the cavern. The advantage of this mode is that it produces saturated brine at an early stage of cavern development. A disadvantage is that the cavern primarily develops at the top, where the brine is dilute. Due to temperature changes in the well bore, moreover, crystallization of salt or gypsum in the supersaturated brine may cause clogging of the tubulars.

In the direct circulation or 'bottom injection' method, fresh water is injected at the bottom, and saturated brine is produced from the upper part of the cavern. This method results in a cavern, in which the brine produced is somewhat diluted by the injection buoyancy plume. Thus, clogging of tubulars is prevented. Generally, mining starts using this direct circulation mode, changing to the indirect circulation mode at some later point in time.

The final shape of a cavern is controlled by the depths of the injection and production tubulars, the mode of operation, the water throughput, the specific dissolution rate of any salt layer, and the setting depth of the blanket fluid. Cavern development preferentially takes place in an upward vertical direction as fresh water is lighter (1000 kg/m³) than saturated brine (1200 kg/m³). In order to allow controlled vertical and lateral development of a cavern, a so-called blanket fluid or gas may be injected (Fig. 9). This fluid or gas should not dissolve salt nor mix with brine, and it should float on brine. Commonly, gasoline is applied (800 kg/m³), but compressed air or nitrogen can be used as well.

Cavern lifetime amounts to tens of years. Sequential cavern development, i.e. the leaching of the void by applying subsequent leaching steps, is simulated and subsequently monitored by time-lapse acoustic measurements. The measurements are acquired using a wireline-mounted probe, which is lowered into the well bore. Comparison with data of former surveys or with results of computer simulations enables the verification of compliance with planned cavern development.

The integrity and long-term stability of a cavern are important issues with respect to potential surface subsidence. Maximum allowable cavern dimensions are determined by finite-element modelling. The mechanical behaviour of the salt is described by means of a constitutive law, which gives the relation between stress and deformation and which incorporates the visco-plastic, time-dependent, non-linear behaviour of salt. Salt behaves like

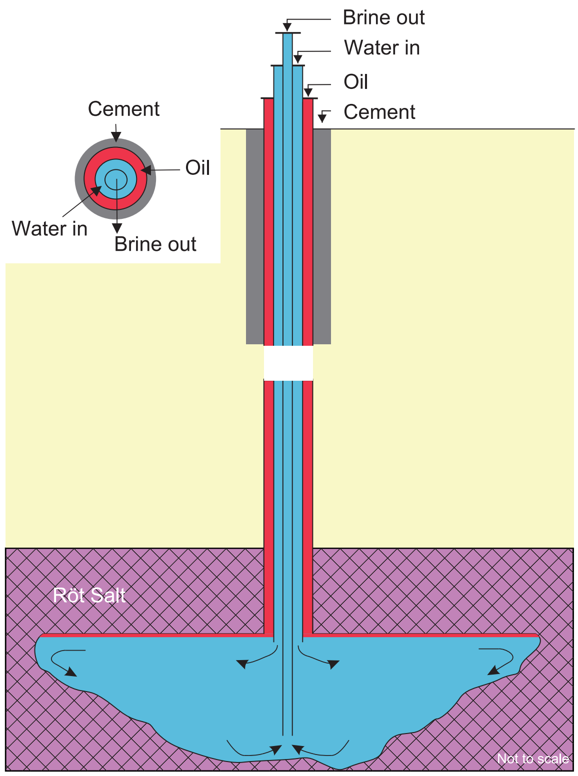

Fig. 9. Cross section of a cavern at a depth of ca. 300 m in the bedded Triassic Röt Salt in the Twenthe-Rijn concession. The flat top of the cavern results from the application of a blanket fluid, and maintains the stability of the roof of the cavern (after Wassmann & Brouwer, 1987).

a viscous fluid upon application of an external stress field. The resulting 'creep' is temperature (i.e. depth) dependent and has a large impact on cavern operations. Over time, creep results in a homogeneous stress field around a cavern, which is a prerequisite for cavern integrity and long-term stability.

Brine production is closely monitored by mass balancing and chemical analysis. In addition, parameters like flow, pressure, temperature, saturation, etc. are recorded continuously to provide information on the actual development of a cavern and on any deviation from default values.

Periodically, the surface subsidence is measured to assure compliance with planned cavern development.

Shallow depth (300-500 m), Twenthe-Rijn

Since production started in 1936, over 450 wells have been drilled and over 70 Mt of rock salt have been mined. Initially, wells were drilled in the direct vicinity of the evaporation plant, south of the city of Hengelo. The mining method evolved over time; at present salt is extracted from caverns which each have three wells at a spacing of 40 m. In the future, caverns with single wells will be developed. Cavern development starts at the base of the salt. Once the initial cavern has developed, production is increased and the cavern is further developed, laterally and vertically, into an ellipsoidal or cylindrical void. The height of a cavern depends on the depth of the salt deposit (more specifically on the distance between the top of the salt and the base of the Tertiary) and the bulking factor, i.e. the ratio between loose and consolidated rock material in the overburden. Caverns are spaced such, that between caverns a safety pillar of proper size is maintained. Thin dolomitic claystones subdivide the Röt salt into the layers A to D (from base to top), and may hamper cavern development. Salt production takes place from layers A, B and (partially) C.

On top of each cavern a safety roof is maintained in the salt layer C to ensure cavern integrity and stability, and to prevent future surface subsidence. Without a safety roof, like in caverns developed before 1980, mechanical failure of the overlying strata by gravity force may occur. This results in an upward vertical displacement of the cavern, and in surface subsidence of several millimetres per year. Once the collapse zone in the overburden reaches the base of the unconsolidated Tertiary (at 60 to 100 m depth) and depending on the height of the remaining void, severe subsidence, up to the formation of a sinkhole, may occur. The latter actually happened in 1991. By limiting the height of a cavern the formation of a sinkhole is prevented. The migration process may last tens of years. The critical area, i.e. the zone of surface subsidence, is limited to an area that is defined by the angle of draw (45°) relative to the base of the Tertiary (Wassmann, 1980, 1983, 1993).

Moderate depth (600-1600 m), Adolf van Nassau

Solution mining at moderate depth is carried out in the Z2 Salt using one well per cavern. Presently, production is from 11 wells in the Winschoten and from 8 wells in the Zuidwending salt dome. The roofs of the caverns are at depths between 600 and 795 m (Winschoten) and 850 and 875 m (Zuidwending). The actual heights of the caverns vary between 420 and 630 m (Winschoten; Fig. 10) and between 356 and 580 m (Zuidwending). The cavern spacing is ca. 250 m, either in a line (Winschoten) or in a hexagonal (Zuidwending) configuration. Cumulative production is over 80 Mt of rock salt since the beginning of the production in the 1950s (Winschoten) and 1960s (Zuidwending) respectively. Surface subsidence due to cavern convergence amounts to 1 to 2 mm/a. For instance, the total cumulative surface subsidence in the centre of the subsidence bowl is presently around 40 mm for the Zuidwending brine field. Like in the Hengelo brine field, a critical area exists which in this case is defined by a 45° angle of draw relative to the deepest point of a cavern.

Large depth (2500-3000 m), Barradeel

The world's deepest solution mine, operated in the Barradeel concession, feeds a vacuum salt factory at Harlin-

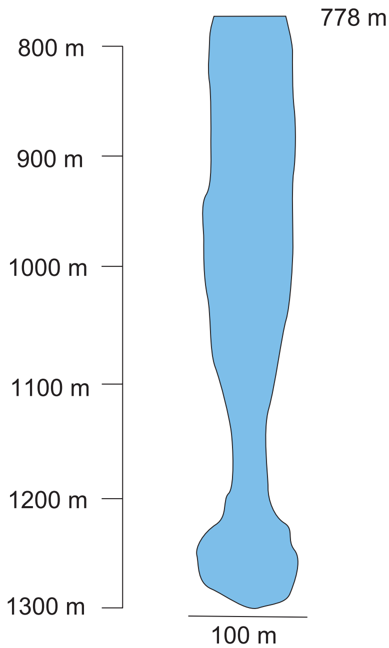

Fig. 10. Cross section of typical cavern in domal salt (Winschoten salt diapir). See Fig. 4 for location.

gen. Two caverns, 500 m apart, are in production since 1995; new caverns at larger distances are under construction. Halite is extracted at 2500-3000 m depth from a 300-m interval in the bedded Z2 Salt. Minor faults affect the base and top of the salt.

The top 30 m of the Z2 Salt is formed by a carnallitic layer. By applying a gasoline blanket the cavern development is kept well below this layer, to avoid carnallite coming in contact with the cavern brine, which would result in contamination of the brine and in loss of the cavern for further production.

Subsurface temperatures at cavern depths are as high as 105° C. As a result, the viscosity of the salt is low, giving a high cavern convergence. This effect is greatly increased by the 28 MPa pressure difference between the cavern pressure (brine hydrostatic of 34 MPa) and the far-field lithostatic pressure (62 MPa) at a depth of 2800 m. The high temperatures do not influence the solubility of NaCl, but significantly increase its rate of dissolution. Where it takes usually one to two years to create sufficient wall surface in a cavern to saturate the brine to acceptable levels for production (ca. 300 kg/m^3), the deep caverns are saturated in one to two months, which is a great advantage for salt production. Although the process of solution mining at large depth is similar to that at moderate depth, the behaviour of caverns strongly differs. At moderate depth, most caverns have a convergence rate that is only 0.05 to 0.1% of the dissolution rate; in the deep caverns, the convergence rate becomes equal to the dissolution rate within a few years. The cavern volume does not grow beyond some 400 000 m^3. The high convergence and the resulting rapid surface subsidence were initially not anticipated due to a lack of experience with deep salt mining worldwide.

The cavern sizes are frequently measured. Although their shapes change somewhat with time, their volumes remain almost constant. The convergence rate, and the subsequent surface subsidence, are directly related to the production rate. The convergence is mainly gravity-driven. The surface subsidence is measured by detailed levelling surveys except for a small part of subsidence bowl in the Waddenzee, where no measurements are possible. Towards the south-east it overlaps with the subsidence area over the gas fields Harlingen and Franeker. The subsidence of Barradeel can be well fitted with a Gaussian equation and amounts to 10 cm/10^6 m^3 of convergence in an area of 6 km diameter. The subsidence bowls generated by both caverns almost fully overlap. At a convergence of 230 000 m^3/a per cavern, the subsidence rate amounts to some 4 to 5 cm/a. A maximum allowed subsidence is determined in view of public concerns about damages, the potential increase of sea-water seepage into the polder area and the subsidence of the sea dike.

Solution mining of potassium-magnesium salt, Veendam

Solution mining for potassium and magnesium salts differs greatly from that of rock salt. They are exclusively mined from bedded salt or salt pillows. In diapirs these salts are strongly folded, and could be extracted only by conventional dry mining. In the Veendam concession, the magnesium component is the main target of mining, while the potassium component is currently not produced.

The mined magnesium salts occur in the Z3 Salt. Three evaporation cycles have been recognized in this salt, and the potassium and magnesium salts are separated by layers of halite. The first, lower, cycle contains the majority of magnesium salts and all of the bischofite; the second and third cycles contain only carnallite. The depth ranges from 1300 to 1800 m (Fig. 11). The layers are dipping about 20° at the mining location. The halite layers occur in all wells with equal thickness and composition, but the magnesium salts are interbedded with thin layers of halite at a centimetre to metre-scale. This interbedding is thought to reflect daily or seasonal variation in the concentration of the depositional brine.

During production the magnesium salts are dissolved preferentially, leaving many halite balconies on the cavern walls. Eventually the balconies will collapse. Sonar surveys, made in the initial stages of cavern development in the 1980s, showed a very irregular cavern shape, but were largely inconclusive in determining the precise shapes and volumes of the caverns due to many reflections from

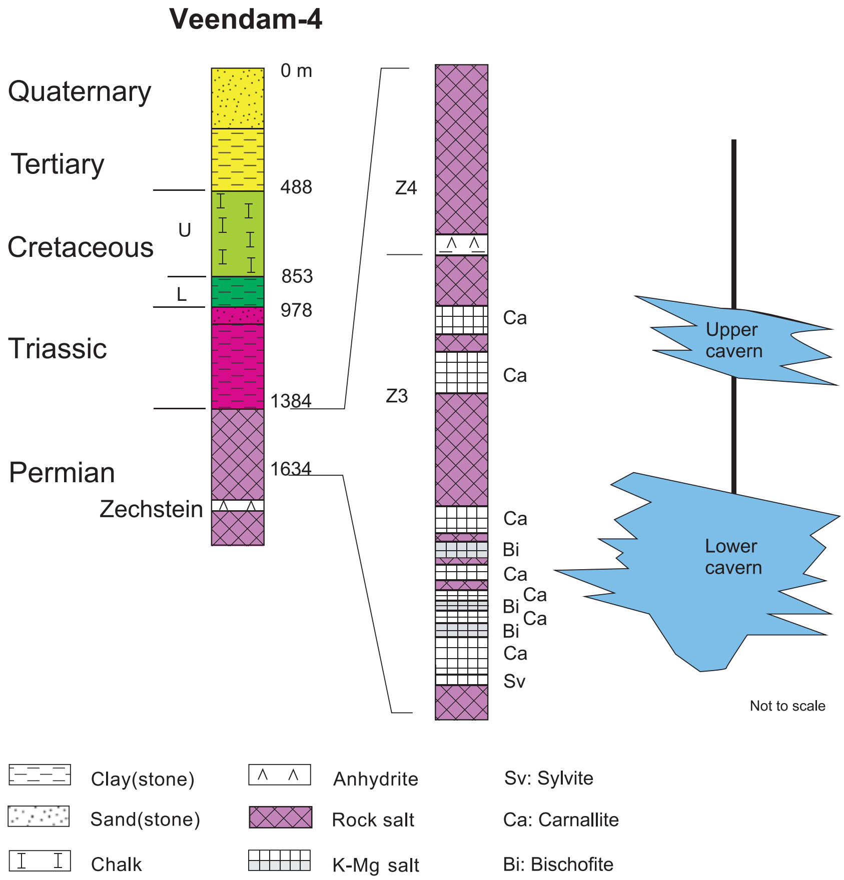

Fig. 11. Stratigraphy and cavern development in the Veendam concession (Veendam-4 well). Cavern shapes reflect dip and solubility of the salt layers. See Fig. 4 for location.

the rock-salt balconies. Presently, the cavern system has a labyrinthine structure, comparable to a natural cave system formed by karstification. Fresh water finds its way to those parts of a cavern that are the easiest to dissolve and the easiest accessible by density flow. The total interconnected area in the cavern field has a diameter of over 1 km. Salt dissolution in a downdip direction is limited, however, to the areas close to the injection points, since there is no mechanism that forces the weak brine down and the saturated brine updip.

The Veendam field contains 12 wells, of which 9 are used for brine production and water injection. Injection and production are intermittently through the same well or via separate wells. Most caverns form part of the labyrinthine multiple-well cavern system, allowing the use of separate injection and production wells.

The (lateral) cavern growth for the same salt-production rate is much larger for bischofite than for halite because: i) the brine, which replaces the dissolved salt contains much more salt (500 kg $MgCl_2/m^3$ versus 300 kg $NaCl/m^3$), ii) bischofite contains more water than solids (850 kg H_2O/m^3 and 750 kg $MgCl_2/m^3$) whereas rock salt contains 2168 kg $NaCl/m^3$, and iii) the cumulative thickness of the Mg-salt layers is ca. 40 to 50 m, compared to hundreds of metres in rock salt.

Fresh water is injected at 30 to 60 m below a rock-salt layer into a cavern, filled with fully saturated brine (1350 kg/m^3 at bischofite saturation). The buoyancy effects make the water rise in the cavern in a turbulent plume, which causes intense mixing with the existing brine. The brine–water mixture is almost saturated with $MgCl_2$ when it reaches the rock salt and can only dissolve negligible amounts of NaCl. The rock salt hence acts as a natural barrier to solution mining. The slightly $MgCl_2$-undersaturated brine will flow, usually following the stratification updip to the highest spot in the cavern, where bischofite or carnallite will be dissolved. Subsequently, the saturated brine will flow down-dip to the deepest point of the cavern, where it can be produced.

If the salts were immobile, or if the cavern pressure could be kept at almost lithostatic, the caverns would continue to grow. However, the magnesium salts have a low viscosity and they flow towards the caverns, driven by the weight of the overburden. Since the magnesium salts occur in layers, salt movement occurs mainly in and through these layers. The salts are squeezed towards the caverns, comparable to tooth paste in a tube or fresh cement between two bricks. This rate of salt flow is such that the brine volume of the total cavern field remains (almost) constant during production: the cavern growth by dissolution is balanced by the cavern convergence.

To monitor subsidence, a dense grid of levelling points is positioned in the area of the subsidence bowl. The bowl shape and rate of subsidence due to salt mining are very different from the overlapping subsidence bowls caused by gas production from the Groningen and Annerveen fields. A slightly oval-shaped bowl, with a diameter of about 3 km, reflecting the non-symmetric layout of the cavern field, is measured. Finite-element-modelling calculations indicate that the surface subsidence closely fits a Gaussian subsidence bowl which mainly depends on the cavern convergence, and only to a minor extent on the time to achieve this convergence (Fokker, 1995). The subsidence at the surface results from several caverns. Since in elastic deformation it is possible to superimpose effects from different sources, and the subsidence is mainly a transmission of local (cavern) effects through an elastic overburden, it is possible to determine the combined effect of more than one squeezing well by superposition of effects from single caverns (Fokker et al., 2000). The measured and calculated subsidence are in the order of 2 cm/a and give an excellent fit throughout the years, resulting in good confidence in the subsidence predictions (Kruse, 1999).

Economic aspects

Approximately 70% of the Dutch NaCl production of 5.5 Mt/a is used as feedstock for the chemical industry, i.e. production of chlorine which is indispensable for the manufacturing of PVC, textiles, aluminium, soap and detergents. Other uses of salt include applications in food and beverages, water treatment, licks for livestock, tanneries and road de-icing. Apart from chemical pureness and constant high quality, the cost of transportation is an important issue for most end-users. This is the reason why salt plants usually have direct access to navigable water.

About 250 000 t/a of $MgCl_2$ is produced, of which 75% is used as brine for the production of high-grade magnesium oxide. The remainder is sold as brine or salt, and transported by ship or by road. The magnesium oxide is sold and used for the production of heat-resistant bricks, which in turn are used in iron and cement furnaces. The brine is used in many applications in the cement industry, but also as de-dusting and de-icing material.

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Groundwater

J.J. de Vries

ABSTRACT

The subsurface of the Netherlands is, from a hydrogeological viewpoint, dominated by a regional aquifer, consisting of medium-grained Plio-Pleistocene fluvial sand with a thickness ranging from 25 to 250 m. The aquifer is at the surface in the eastern half of the country and dips below semi-confining layers of lagoonal clay and peat in the western coastal area. The groundwater table is close to the surface almost everywhere and the precipitation surplus of about 300 mm/a is discharged by a dense and largely artificial drainage system. Most of the western half of the country is below sea level and consists of a patchwork of polders, each with its own artificially controlled level of surface water and groundwater. This has resulted in a complex system of infiltration into relatively elevated polders and in discharge by diffuse upward leakage of fresh and brackish water into deep polders. The eastern half of the country comprises (i) shallow aquifers in level areas, which are drained by seasonally contracting and expanding stream systems, and (ii) deep aquifers in more elevated areas with hardly any surface drainage and a groundwater table that reacts predominantly on annual variations in rainfall. A total volume of the order of 1700 million m^3 of fresh groundwater is annually extracted in the Netherlands; about 60% of this is used for public water supply.

Keywords: Netherlands, hydrochemistry, hydrogeology, coastal aquifer, groundwater management

Introduction

Groundwater occurrence

Groundwater is present in the pores and fractures of the entire sedimentary sequence of the onshore and offshore Netherlands. The geological framework of the subsurface controls the spatial distribution of porosity and permeability, and thus the distribution, thickness and structure of the hydrostratigraphic units.

Flow in the upper groundwater zone is primarily driven by rainfall-induced and topographically controlled, potential energy. At greater depths, groundwater becomes increasingly separated from the present-day hydrological cycle. In general, there is a gradual transition between the shallow and deep zones, depending on the occurrence and character of the poorly-permeable layers in between the two zones, and the persistence of the, either natural or artificially induced, vertical flow components. The turnover time of water circulation in the upper zone is of the order of days to thousands of years (locally more than 10 000 years), whereas water in the lower zone circulates on geological time scales.

The actual hydrogeological conditions in the lower zone are strongly influenced by residual components from the complex geological history. These deep groundwater systems are mainly driven by pressure gradients, caused by large-scale (paleo-)topography and tectonic forces, and by thermo-chemical processes. Permeable formations in the deep zone consist mainly of sand, sandstone and chalk. The low-permeable units include layers of clay, claystones and particularly the evaporites of the Permian Zechstein Group and to a lesser extent the Upper Germanic Trias Group (Verweij, 2003). The deformation history and the associated deep faults and salt structures disrupt the lateral continuity of the permeable and low-permeable units.

The upper groundwater systems consist of components of different orders of magnitude of regional extent and depth of penetration, which are genetically related to the land-surface topography and the associated water-table topography (Figs 1, 2). The main elements of the topographic relief within the Dutch landscape are the Holocene coastal dunes, the remnants of glacial features of Pleistocene origin in the central-eastern part of the country, and the pre-Pleistocene uplifted parts of the southernmost Netherlands (Fig. 1). The penetration depth of the (first-order) topography-driven groundwater systems is only in the order of hundreds of metres due to the presence of low-permeable clay layers or a succession of low-permeable sediments of mainly Early Pleistocene and older age. The maximum depth is reached in the Roer Valley Graben (cf. Fig. 6), where the influence of a supra-regional flow system is observed between depths of 500 and 1000 m below surface.

There is increasing awareness that groundwater plays an important role in geological processes. Understanding of the hydrogeological conditions is important for the exploration and exploitation of natural resources such as water, oil, gas, salt, surface mineral deposits and geothermal energy, as well as for the management of the storage of energy carriers and different types of waste, and for the prediction of geohazards. This chapter mainly concerns the upper groundwater systems. For knowledge of the deep groundwater systems in relation to the geological history

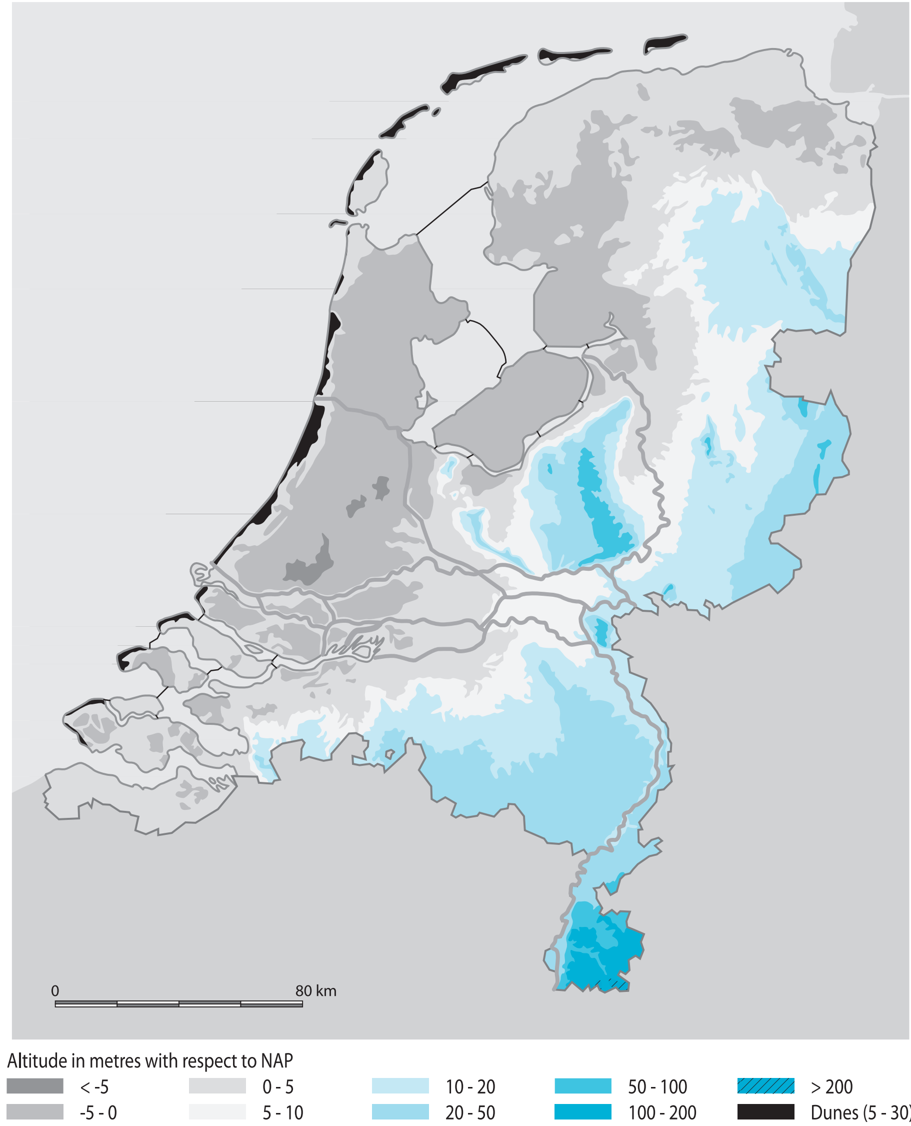

Fig. 1. Surface topography of the Netherlands; elevation relative to NAP = Normaal Amsterdams Peil = Dutch ordnance datum, approximately mean sea level (source: Dufour, 2000).

in the onshore and offshore Netherlands the reader is referred to Verweij (2003).

Development of groundwater research in the Netherlands

Groundwater research in the Netherlands is principally driven by the need to solve the specific water-related problems of coastal lowlands. These problems are in particular related to the search for suitable drinking water and the management of groundwater in an area with brackish groundwater and shallow groundwater tables in polders below sea level. Already during the founding of hydrology as a science in the 19th century, Dutch engineers and geologists, in an attempt to resolve these problems, made fundamental contributions to the concepts that at the beginning of the 20th century resulted in the basic theory of groundwater flow (De Vries, 1982, 2004).

The first systematic groundwater investigations in the Netherlands were carried out in the mid-19th century by Pieter Harting, a physician, geologist and professor of natural history. His interest in groundwater was, apart from scientific curiosity, stimulated by his concern to improve public health by the supply of good drinking water. Because of a lack of basic knowledge of hydraulics, his analysis did not really contribute to a better theoretical understanding of groundwater behaviour, but he nevertheless improved the general knowledge of the occurrence and quality of groundwater in its geological context (see Appendix). A first step in scientific groundwater exploration was made between 1839 and 1844 with the drilling of a

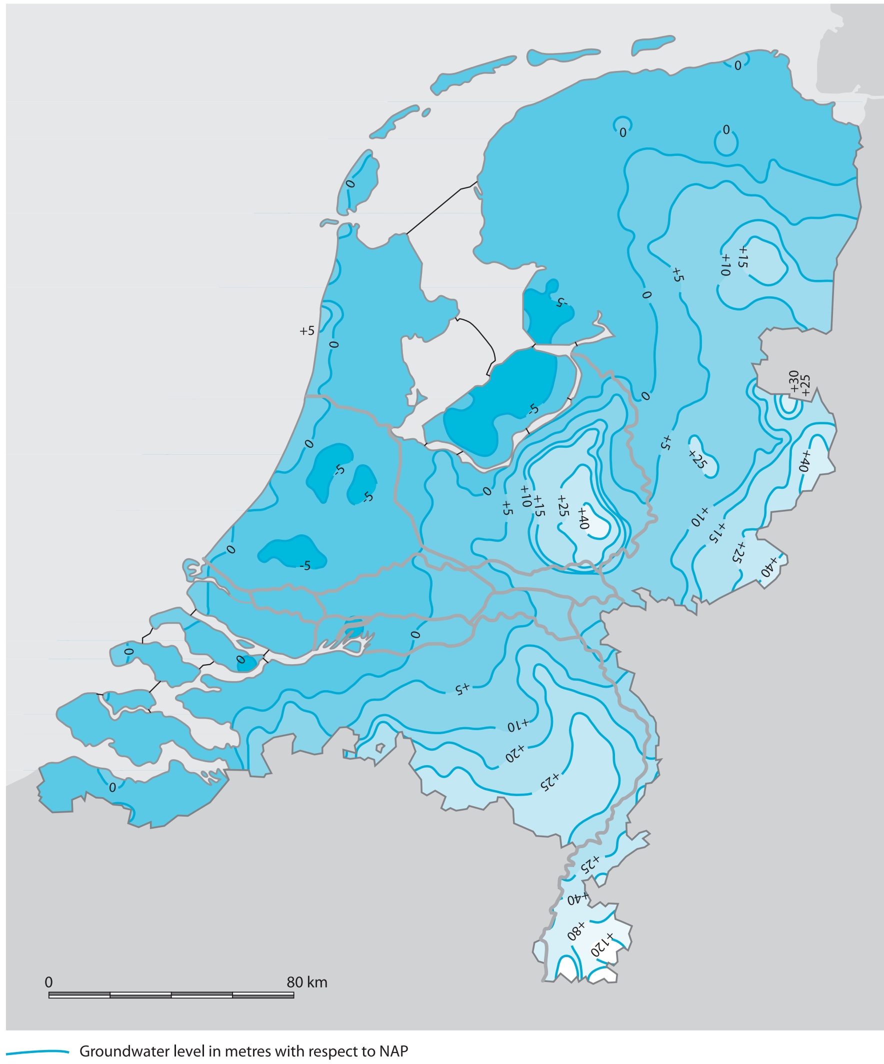

Fig. 2. Groundwater table relative to NAP, observations 9 February 1996 (source: Dufour, 2000).

well to a depth of 172 m at the Amsterdam Noordermarkt. Groundwater quality and yield of this well were disappointing, but the conscientious description and analysis of the soil samples by Harting (1852) constituted an important step towards the knowledge of the subsurface.

After persistent problems with polluted and brackish drinking water, the major towns in the west of the Netherlands eventually went to the groundwater reserves in the coastal dune belt for their water supply. In 1853, Amsterdam was the first to begin with the exploitation of the dune area near Haarlem, 30 km to the west. The main hydrogeological problem was that the small catchment area was surrounded by saline water, thus prone to depletion and salinization. Moreover there was no insight into the vertical extent of the fresh-water reserves, and their origin and replenishment were a matter of wild speculation. Hypotheses about the source of this water varied from local rainfall and condensation to artesian water veins originating in the remote higher grounds of the ice-pushed ridges, or even in the Ardennes in Belgium. Because of the fear of salinization, one prudently started with shallow extraction by drainage canals.

The first Dutch fundamental contribution to groundwater hydrology concerned the position of the interface between fresh and salt water below the dunes. In the 1880s, Captain Willem Badon Ghijben of the Army Corps of Engineers proposed the principle of a fresh-water lens, floating on salt water, which was recharged through local rainfall (Drabbe & Badon Ghijben, 1889). According to this principle, the thickness of the fresh-water pocket below the higher dunes along the Dutch coast was predicted to be not less than 150 m on average (Fig. 3). This remarkable hypothesis remained unnoticed until the German engineer Herzberg (1901) arrived independently at the same concept, which is now well-known as the Ghijben-Herzberg principle. Although the existence of this extensive fresh-water pocket was indeed proven by exploration drillings at the turn of the century, it took another 20 years before it was generally accepted that this fresh water originated from the limited input by local rainfall and not from an inexhaustible artesian inflow.

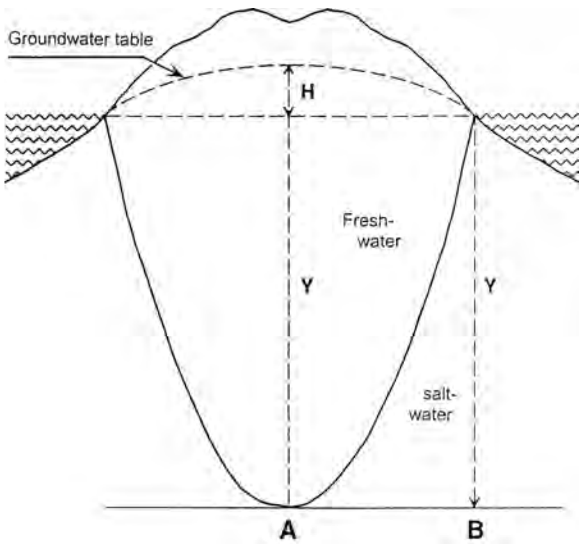

Fig. 3. The position of the fresh–salt water interface according to the Ghijben-Herzberg principle. The weight of the column of fresh water at A: $(y + H)\rho_f$, equals the weight of the column of salt water at B: $(y \cdot \rho_s)$, where ρ_f and ρ_s are the specific weights of fresh water (1 g/cm^3) and salt water (1.0238 g/cm^3) respectively, so that $y = 42H$. The average maximum height H of the water table in the Netherlands coastal dune area is in the order of 5 m above sea level.

At the turn of the century, Johan M.K. Pennink, director of the Amsterdam Municipal Water Works, warned against over-exploitation and salinization of the dune catchment. To convince the municipality, he simulated the upward movement of the salt-water boundary by overpumping, through experiments with viscous parallel-plate models. In order to cope with the growing water demand, he proposed artificial recharge of the dune area by river water, for which he developed a detailed scheme. His plans were only implemented half a century later once salinization problems had become serious. The onset of the salinization process was convincingly demonstrated by Pennink in a sound report published in 1914. Pennink experienced a long-lasting conflict with the Amsterdam municipality because he stubbornly refused to extract more water from the dunes than a percentage of the quantity he rightly assumed to be the rainfall replenishment.

Other even more essential contributions by Pennink were his field experiments to investigate the groundwater flow pattern around the drainage canals. Until then it was generally assumed that groundwater under free water-table conditions could not move in an upward direction, so that the flow to a drainage canal was limited to the depth of the canal. Pennink measured the distribution of the hydraulic head with a row of piezometers at different depths in a section perpendicular to the canals. By drawing the flow lines perpendicular to the measured equipotential surfaces, he proved the radially upward converging flow pattern near the canals, which explained the observed increase in hydraulic head with depth below the discharge areas. This downward increase in hydraulic head was pre-

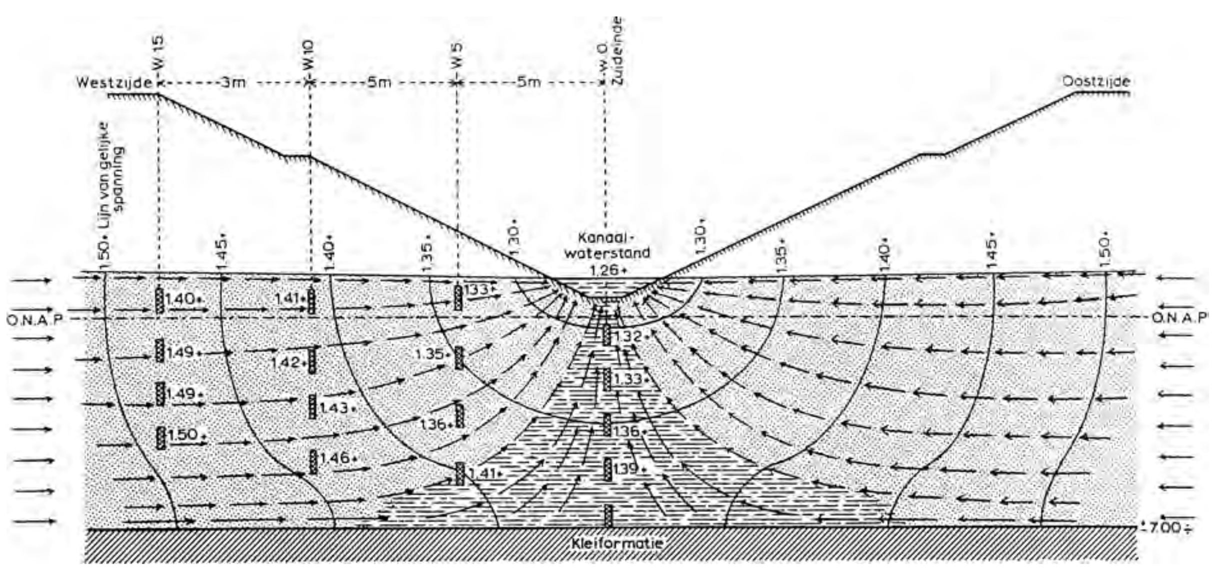

Fig. 4. Flow lines around a drainage canal based on hydraulic-head observations in piezometers; redrawn from Pennink (1905); kanaal-waterstand = water level in canal; lijn van gelijke spanning = line of equal hydraulic head; kleiformatie = clay layer.

viously used as an argument in favour of the existence of artesian water (Pennink, 1905; Fig. 4).

In fact, Pennink solved by performing field experiments, the continuity equation and made clear that groundwater flow is governed by a combination of the flow equation according to Darcy and the continuity equation according to Laplace. At about the same time, physicists like Ph. Forchheimer, C.S. Slichter and J. Boussinesq, proposed this concept on theoretical grounds. Pennink was not familiar with this theoretical approach, but he certainly was the first to prove experimentally the validity of this concept, that eventually at the beginning of the 20th century resulted in a general theory of groundwater flow.

This theory made it possible to simulate groundwater flow in a mathematical model, and to solve flow problems as boundary-value problems. The first Dutch contribution along this line was the solution of the typical lowland problem of groundwater flow in a leaky aquifer with abundant water at the surface. The derived formulas enabled the quantitative description of the complicated groundwater-flow conditions in polder areas and the prediction of the consequences of, for instance, groundwater extraction, water management and land-reclamation works. As early as 1914, J. Kooper, captain in the Army Corps of Engineers and engineer with the National Bureau for Drinking Water Supply (established in 1913), published the solution for the flow to a well in a leaky aquifer. His flow equations were subsequently elaborated by G.J. de Glee in 1930, and his formula for flow to an extraction well has become generally known as the 'De Glee formula' (Fig. 5). This solution appeared in the international literature only after World War II through the work of C.E. Jacob and M.S. Hantush in the USA.

A scientific approach to the problem of land drainage emerged in the 1930s under the leadership of S.B. Hooghoudt at the Experimental Station and Soil Science Institute in Groningen. These trail-blazing studies of Hooghoudt and his collaborators, notably L.F. Ernst, which combined theoretical analyses and plot-scale experiments, have led to a basic understanding of the processes of groundwater drainage and a sound basis for groundwater-table management (Hooghoudt, 1940; Ernst, 1962). Hooghoudt's program was relocated to the Institute for Land and Water Management Research (ICW) in Wageningen in the 1950s, which meant a shift of research focus to the influence of groundwater depth on capillary transport processes in the soil-water zone and the associated evapotranspiration and crop production.

Until the end of World War II, groundwater research and its application in water management in the Netherlands developed mainly along sectoral lines with separate solutions for problems related to public water supply, agriculture and general water management, including protection against flooding and salt encroachment. After the

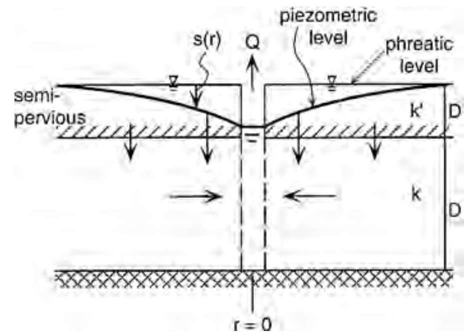

Fig. 5. Steady-state radial symmetrical flow to a well in a leaky aquifer, according to Kooper (1914) and De Glee (1930). For a small-diameter well in an infinite flow field, the solution is: $s = \frac{Q}{2\pi T} K_0\left(\frac{r}{\lambda}\right)$, with $\lambda = \sqrt{Tc}$, and where *s* is the lowering of hydraulic head at distance *r* from the well through a groundwater extraction *Q*; *T* and λ are transmissivity and leakage factor of the aquifer respectively; *c* is the vertical flow resistance, which equals D'/k', where D' and k' are the thickness and vertical hydraulic conductivity of the semi-pervious confining layer; K_0 is a modified Bessel function of zero order. This equation, and related formulas for flow around large-diameter excavations or polders, have been widely used for pumping-test analyses, evaluation of the impact of groundwater extractions, and assessments of groundwater inflow into deep polders and excavations.

war, it became clear that an integrated approach was required to deal adequately with conflicting interests. This led to the founding of the Committee of Hydrological Research (CHO-TNO), under the umbrella of the Netherlands Organisation for Applied Scientific Research TNO. This Committee has played an eminent role in stimulation and coordination of hydrological research until the early 1990s. It perfectly documented the hydrological activities and achievements during this post-war period in a series of technical reports and proceedings of meetings. Another post-war initiative was the establishment of an archive for groundwater levels, which were systematically monitored in thousands of observation wells with the help of hundreds of volunteers. This work got a great impulse by a national inventory of the agricultural water condition by the Committee on Agro-hydrological Research (COLN-TNO), which was predominantly based on a survey of the groundwater-depth regime by a network of 23 000 observation wells (one per 100 ha) during the period 1952-1956 (Visser, 1958).

The spirit of cooperation led to integrated regional surveys, which not only focussed on a more balanced water management, but also encouraged fundamental research on the interaction of surface water, groundwater, soil water and evapotranspiration. The thorough investigations in the small catchment of the Leerinkbeek in east Gelderland (Colenbrander et al., 1970) are an early example of this approach. The awareness of environmental pollution subsequently stimulated the study of hydrogeochemical aspects of groundwater. This resulted, for instance, in a textbook on hydrochemistry and groundwater pollution by Appelo & Postma (1993). To sustain the regional and integrated water management, a national groundwater-mapping, monitoring and data-management programme was established under the umbrella of TNO in the late 1960s. These groundwater surveys and analyses were based in particular on an approach in which coherent groundwater systems are considered as the basic units. The hydrogeological investigations gradually included other geotechnical aspects as well. This eventually resulted in the 1990s in a concentration of all national groundwater and geological surveys within the Netherlands Institute of Applied Geoscience (TNO-NITG).

The extensive contributions of Dutch hydrologists to the theoretical solution of groundwater-flow problems are presented in several internationally recognized textbooks, including Verruijt (1970, 1982), Kruseman & De Ridder (1970, 1990), Huisman (1972), Huisman & Olsthoorn (1983) and Bruggeman (1999).

Hydrogeological setting

Geological framework

Almost the entire territory of the Netherlands is part of the south-eastern marginal zone of the subsiding North Sea Basin. The limits of this basin are close to the south-eastern and eastern national boundaries, where they form a transition to the relatively more stable and uplifted areas of the Ardennes and the Rhenish Massif. The average elevation at the southern and eastern basin boundaries is normally of the order of 30 to 40 m, while the maximum elevation in the Netherlands of 322 m above NAP is at the southernmost tip of the country (NAP = Normaal Amsterdams Peil = Dutch ordnance datum = approximately mean sea level). Other relatively high areas are the Pleistocene ice-pushed hills in the central-eastern part of the country, notably the Utrechtse Heuvelrug and the Veluwe, with a maximum elevation of 107 m above NAP (Figs 1, 6).

The deposits participating in the present-day hydrological cycle, consist predominantly of Plio-Pleistocene, medium to coarse, fluvial sands with a thickness that increases north-westward to more than 300 m (Fig. 6; Breeuwer & Jelgersma, 1973). These sediments belong to the Upper North Sea Group and include the Echteld, Kreftenheye, Urk, Sterksel, Waalre, Beegden, Appelscha, Peize and Kieseloolite formations. The Appelscha and Peize formations were deposited by the former Eridanos River, which originated in the Scandinavian-Baltic area; the other formations have a Rhine and Meuse origin (De Mulder et al., 2003; De Gans, this volume). The lower part of the aquifer partly consists of a succession of coarse and fine, marine sediments of the Early Pleistocene Maassluis Formation and the Pliocene Oosterhout Formation.

The aquifer sands have an average permeability factor that ranges from 20 to 50 m/day, so that the average transmissivity of the subsurface is several thousands to locally more than 10 000 m^2/day (for explanation of these parameters, see Appendix). Semi-confining layers of various extent cause the Plio-Pleistocene fluvial deposits to behave as a multi-layer aquifer. On a regional scale, however, notably in the western half of the country, the aquifer can roughly be divided in an upper and a lower aquifer, separated at about 50 m depth below sea level by Middle Pleistocene clayey deposits. Glacial and fluvio-glacial clays of the Peelo (Elsterian) and Drente (Saalian) formations form extensive aquicludes in the north and north-east (Figs 7a, b). The Plio-Pleistocene aquifer is the most important source for public water supply in the east and south. Groundwater in the Pleistocene aquifer in the west is predominantly brackish due to marine influences during the Holocene.

The Pleistocene aquifer is at the surface in the east and dips under the clayey and peaty Holocene layers in the coastal area, and thus the upper aquifer shifts from predominantly phreatic in the east and south to semi-confined in the west and north. The Holocene consists mainly of fluvial, tidal-flat and estuarine deposits, accumulated under the influence of a rising sea level behind a series of coastal barriers. These deposits reach a maximum thickness of 25 m near the coast and are partly separated from the sea by a ridge of young coastal dunes, which locally reach a height of 50 m on top of the barrier. The main part of the Holocene coastal lowland is situated below sea level. The Pleistocene aquifer and the Holocene semi-confining layers, with abundant water at and near the surface, are in close communication because of vertical leakage, which is mainly induced by the artificial abstraction of water from the low polders (Figs 1, 2).

The Plio-Pleistocene aquifer lies almost everywhere on a sequence of clayey and sandy deposits of mainly marine origin, ranging in age from Oligocene and Miocene along the southern and eastern fringes, to Pliocene and Early Pleistocene in the western and central part of the basin (Fig. 7a). These low-permeable, basal sediment complexes contain predominantly brackish and saline water and form almost everywhere the lower boundary of the replenished groundwater system.

Unusual groundwater circulation occurs in the SE-NW running Roer Valley Graben system that cuts into the south-eastern basin margins (Figs 6, 8). Here the Plio-Pleistocene fluvial deposits are underlain by more than 1500 m of fine-grained sands and clays of mainly marine origin and Miocene and Late Oligocene age (Breda and Veldhoven formations). The salt–fresh water interface in the Breda Formation reaches depths of about 1000 m near the German border, which indicates that the marine sediments in the Roer Valley Graben have been desalinized by fresh groundwater inflow from the past and present outcrops of Tertiary sands in the east and south.

A series of continental sandy aquifers with intercalated lignite horizons (Ville and Inden formations) interdigitates with the Miocene marine clays. The lignite is being mined in Germany in open pits of hundreds of metres depth, just across the Netherlands-German border. Groundwater extraction to drain the excavations has caused a decline of more than 10 m in hydraulic head in some parts of the deep confined aquifers in the Dutch section of the graben, at a distance of more than 50 km from the excavations (cf. Fig. 16). Groundwater ages in this deep and supra-regional aquifer system range from 1000 years near the Dutch-German border to more than 10 000 years some 50 km to the north-west (Stuurman, 2000).

Other pre-Plio-Pleistocene aquifers are present in the uplifted area south of the Feldbiss Fault system, which forms the southern boundary of the Roer Valley Graben. Aquifers occur in Miocene fine-grained sandy layers in the northern part of this uplifted area, and in Lower Paleocene and Upper Cretaceous karstified chalk and marls further to the south. Both aquifers have a thickness in the order of 100 m. The Miocene aquifer is partly covered by Plio-Pleistocene gravel, sand and clay, whereas the northern part of the Cretaceous-Paleocene aquifer is overlain by confining Oligocene clay. The Oligocene clay dips to the north to form the lower boundary of the Miocene aquifer. The chalk aquifer is underlain by Upper Cretaceous sandstone, sand and clay of the Aken and Vaals formations (Fig. 8; Patijn, 1966; NITG, 1999).

The uplifted area, south of the Feldbiss Fault, forms a plateau, which is drained by incised branches of the river Meuse, notably the Geul in the south and Geleenbeek in the north. The exposed chalk area has a typical karst appearance with bowl-shaped depressions and dry valleys. The main draining river, the Geul with its tributary the Gulp, receives water from a number of springs, which particularly emerge in upstream areas where the river bed cuts into the interface of the chalk and the impervious clay. Several of these springs, producing tens of cubic metres per hour, have been intercepted and encased for drinking-water supply (Waterleiding Zuid-Limburg, 1941). The chalk is an important source for this supply, yielding up to 6 million m^3/a by the best producing group of wells. Large extractions take place along the Heerlerheide–Benzenrade fault system, where obstruction of groundwater flow by low-permeable fractures has resulted in concentrated vertical fluxes with associated chalk dissolution and enhanced permeability.

Groundwater at greater depths, separated from these recharged aquifers by low-permeable clayey deposits occurs in sandstone and quartzite inter-beds in Carboniferous shales and claystones. The main conduits through these low-permeable beds are related to NW-SE running fractures. In contrast, NE-SW directed fractures create barriers to groundwater flow. Breaches through these fracture aquifers during coal mining often produced temporary yields of hundreds of cubic metres of water per hour. Accumulated extraction for dewatering until the end of the mining activities in 1974 totalled 25 million m^3 (NITG, 1999).

Fig. 6. Approximate thickness in metres of that part of the Plio-Pleistocene aquifer that is involved in the present-day groundwater circulation (after De Vries, 1974); map also shows topographic names referred to in text; S and B are locations of the polders (land reclaimed from lakes) Schermer and Beemster.

There is a chemical stratification, ranging from $Ca(HCO_3)_2$ dominated water in the Cretaceous chalk, to NaCl water at depths below 500 m in the Carboniferous sediments. In between is $NaHCO_3$ water, formed by cation exchange during freshening of the original salt water (cf. section 'Process reconstruction and prediction'). Thermo-saline water with temperatures up to 50°C, ascends locally along deep fractures (Kimpe, 1963; NITG, 1999).

Most of the information on the subsurface and its

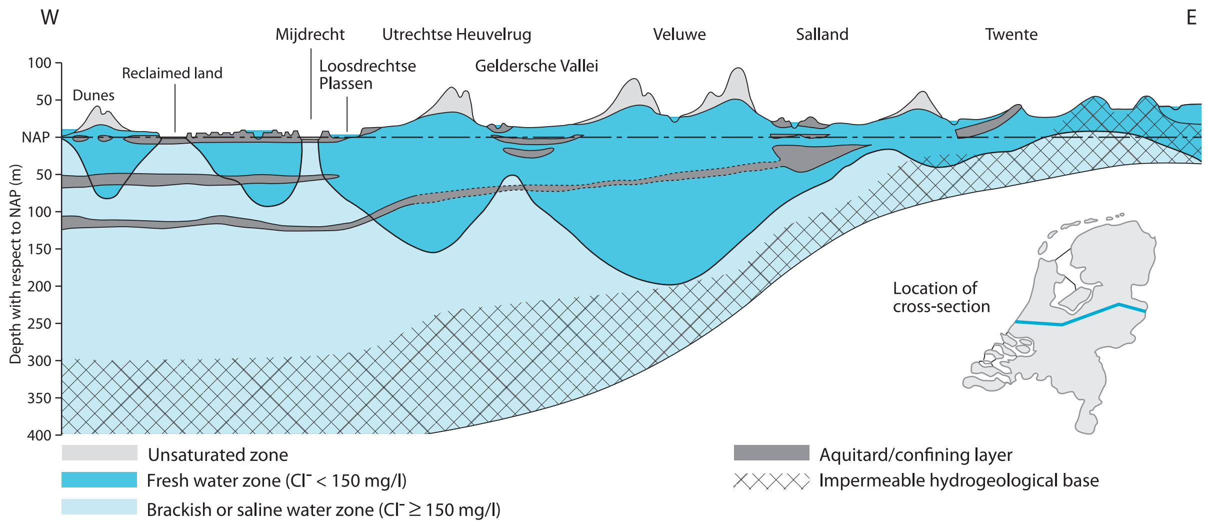

Fig. 7a. Schematic topographic-hydrogeologic east-west section showing the approximate depth of the fresh–brackish water interface (brackish > 150 mg Cl/l); length of section ca. 200 km (after Van de Ven (ed.), 1993; source Dufour, 2000).

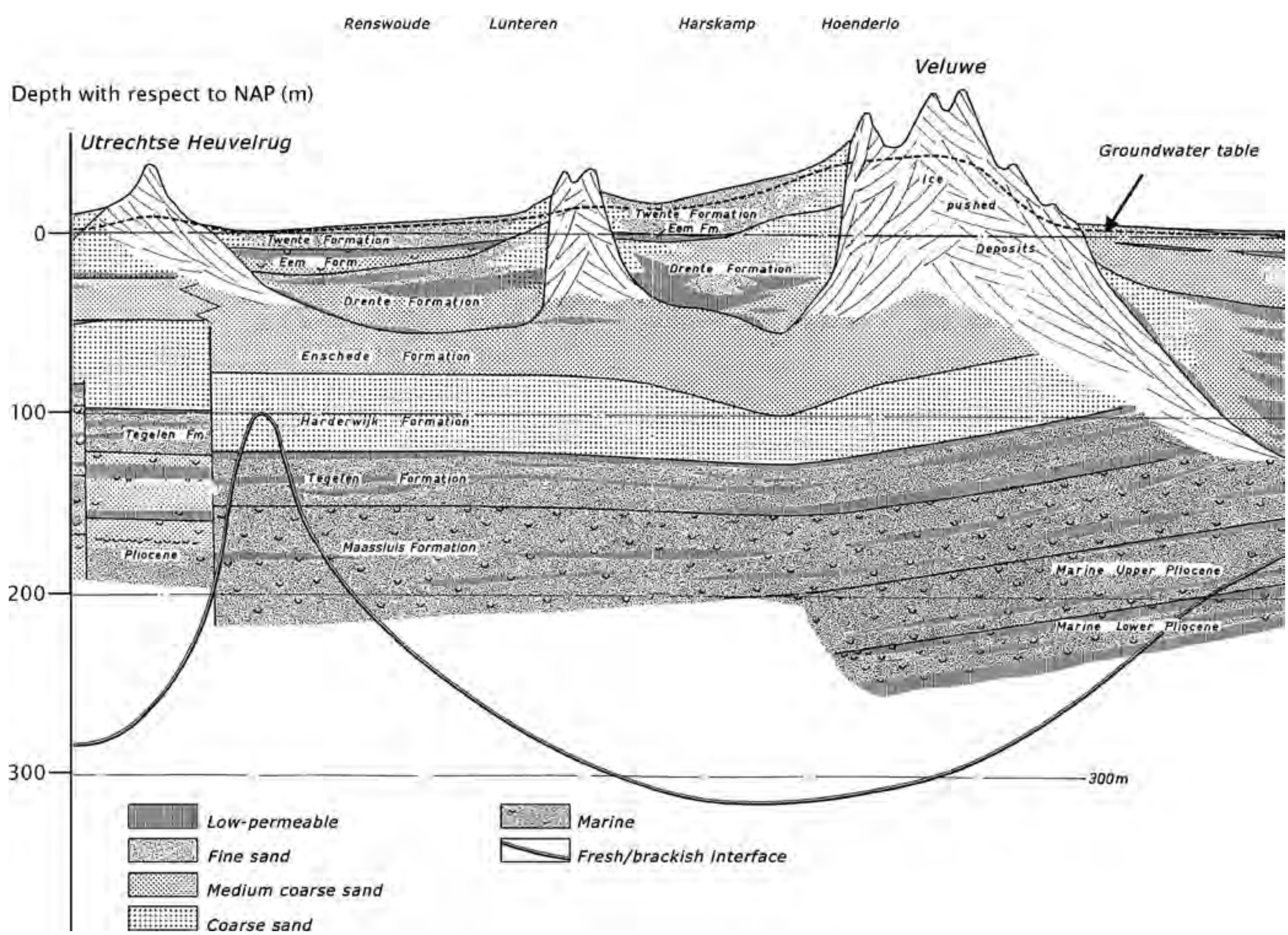

Fig. 7b. Detail section, WSW-ENE and 12 km south of Fig. 7a (after Breeuwer & Jelgersma, 1973, with stratigraphic nomenclature given by these authors). Length of section is ca. 60 km; the deeper position of the fresh–brackish interface in comparison with Fig. 7a is due to depth increase of the interface in south-eastern direction.

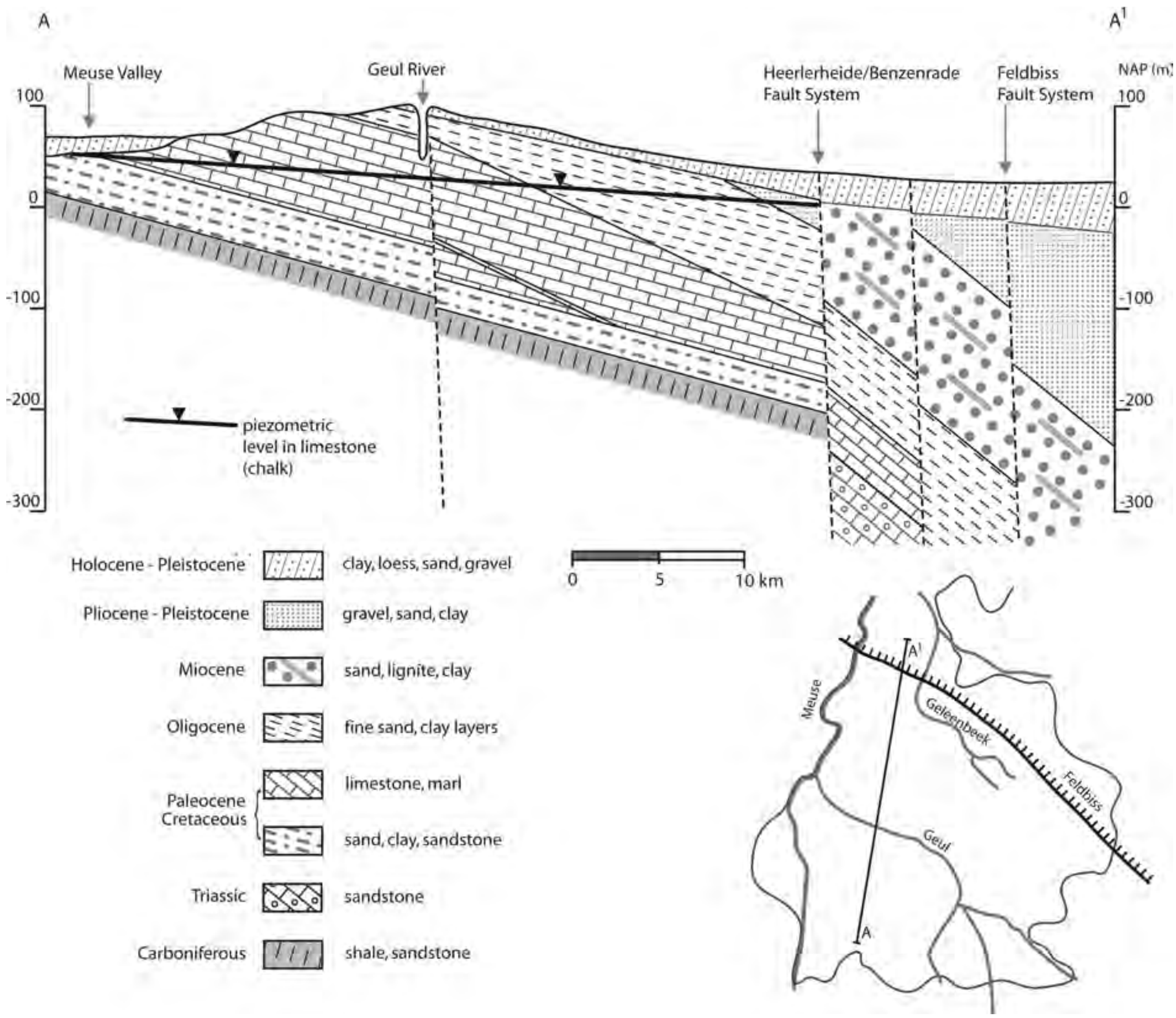

Fig. 8. Schematic hydrogeological section through the Cretaceous-Paleocene chalk aquifer in Zuid-Limburg (redrawn from Patijn, 1966). The piezometric level indicated refers exclusively to the chalk aquifer.

groundwater is contained in the archives of TNO-NITG and has been compiled in (digital) maps and databases. For a comprehensive overview of the groundwater situation in the Netherlands, the reader is referred to Dufour (1998, 2000).

Climate and groundwater recharge

Annual rainfall and evapotranspiration in the Netherlands are of the order of 800 and 500 mm respectively. Rainfall is evenly distributed over the seasons, but evapotranspiration is highest during summer, when it is more or less in balance with precipitation. Thus the average annual precipitation surplus of 250 to 350 mm is concentrated in the winter period (e.g. Meinardi, 1994). The rainfall surplus can be much higher in sandy areas with little or no vegetation; annual values of more than 600 mm were observed in bare dune sand (Stuyfzand, 1993). The lowest values, less than 150 mm, were measured in pine forest, where high interception-evaporation reduces infiltration (Stuyfzand, 1993; Gehrels, 1999). Inflow of water through the Rhine and Meuse amounts to an extra equivalent supply of 2100 mm/a (78 000 million m^3/a) on average. This river water is mainly used to combat salinization in the coastal area and to supplement public, industrial and agricultural water supply (NHV, 2004).

Almost all rainfall infiltrates the subsurface, and that part that does not return to the atmosphere by evapotranspiration, is subsequently discharged as groundwater by a dense and in many places artificial drainage system. The drainage capacity of this system normally increases with decreasing groundwater depth and reaches a maximum of the order of 15 mm/day in relatively low-lying areas with shallow groundwater tables and a dense drainage network. This applies particularly to the coastal lowlands where groundwater levels are artificially controlled by pumping. The clay and peat-containing soils in this area are strongly anisotropic due to low-permeable horizontal sediment layers. This obstructs vertical penetration of the precipitation surplus and stimulates concentrated lateral drainage through the upper metres. The marine Holocene deposits are extensively freshened, but pockets of remnant brackish water in low-permeable strata indicate the preferential flow of fresh water through relatively permeable sediments in buried tidal channels.

The groundwater table is shallow and almost everywhere less than 2 m below the surface with a maximum seasonal fluctuation of about 1 m, except for ice-pushed hills in the central and eastern Netherlands and the uplifted areas in Zuid-Limburg, where groundwater tables are much deeper. Figure 2 gives the overall pattern of the hydraulic head in the Pleistocene aquifer. Superimposed on this general pattern is the higher-order groundwater-table topography, related to the drainage system of small streams and ditches.

The Holocene coastal lowland

Hydrogeological evolution

The sea has repeatedly invaded the western part of the Netherlands during the Quaternary. Figure 9 shows the most easterly positions of the Pleistocene and Holocene coastlines and the overall present-day depth of the fresh–salt water interface. Saline groundwater is almost everywhere present in the marine Pleistocene and Tertiary deposits and has intruded a large part of the overlying fluvial Pleistocene aquifer in the coastal area. The maximum depths of this fresh–salt interface are reached below the coastal dunes and the ice-pushed hills, where the relatively elevated groundwater tables caused deep infiltration of fresh meteoric water (Figs 7, 9). Other deep fresh-water occurrences are found in the Tertiary aquifers in the south-east, where subsurface inflow from the outcrops situated further to the east and south, has flushed the marine sediments by (supra-)regional groundwater flow (cf. 'Geological framework').

Invasion by the sea took place mainly in four periods after the Early Pleistocene: Cromerian, Holsteinian, Eemian and Holocene. In most areas, salt water originated from the Pleistocene transgressions has been flushed from the coarse fluvial deposits during the last phase of the Pleistocene, the continental Weichselian, which lasted some 70 000 years. At the beginning of the Holocene a zone of brackish water below an eastward-dipping interface was probably present in the lower part of the aquifer as a result of diffusion, dispersion and compaction-driven flow from the lower marine sediments (Meinardi, 1991; Kooi & De Vries, 1998).

The Holocene sedimentation began with the development of a thick peat layer on the Pleistocene surface as a result of stagnating drainage in front of the encroaching sea. Subsequently, the peat layer became covered by low-permeable lagoonal clayey deposits, which protected the Pleistocene aquifer from salt-water intrusion by density currents, except where they became incised by deep tidal inlets. Recent transport and hydrochemical process modelling and isotope dating, make plausible that large areas may have been salinized as a result of vertical and lateral spreading by free convection of salt water from the large tidal channels (Kooi et al., 1999; Post, 2004).

Illustrative for the process of salinization through a transgression over a protecting clay layer, are the developments in the former Zuiderzee, an embayment of the North Sea, north-east of Amsterdam. This bay was separated from the sea in 1932 by a barrier dike (Afsluitdijk) and turned into a fresh-water lake (IJsselmeer), in which 2000 km^2 of polderland have subsequently been reclaimed (Figs 1, 6). The Zuiderzee came into being during medieval times by destructive encroachment of the sea through tidal inlets in this former peat area. The Holocene clayey deposits at the bottom of the sea protected the underlying Pleistocene aquifer in most of the area from salt-water intrusion by density currents, because there were no deeply incised tidal gullies opening up the aquifer from above. This is evident from Figure 10a which shows a gradual decrease of chloride content in the Holocene layer with depth, from 6000 mg/l (average chloride content of Zuiderzee) to less than 1000 mg/l at the Holocene/Pleistocene boundary. In the Pleistocene aquifer, the chloride content increases again with depth, to more than 6000 mg/l at the lower boundary of the aquifer at a depth of about 200 m (Fig. 10b).

Volker explained the Holocene profile by downward diffusion of salt from the Zuiderzee into the Holocene deposits, and attributed the Pleistocene profile to upward diffusion from the marine Lower Pleistocene into the overlying fluvial Pleistocene deposits (Volker, 1961; Volker & Van der Molen, 1991). However, deep borings in the marine Maassluis and Oosterhout formations revealed salt-water inversions, with lower chloride concentrations in the upper parts of these marine formations below a zone of higher salt content (Fig. 10c). These findings contradict the diffusion hypothesis for the Pleistocene aquifer but are consistent with salinization of this aquifer by density currents during Holocene transgressions. Age determinations of this brackish water in the lower part of the fluvial deposits also suggest the absence of important pre-Holocene components: more than 80% of 275 analysed groundwater samples from the Pleistocene aquifer in the west of the country indicate an age of less than 10 000 years (Post, 2004). Other inversions were encountered in the upper part of the Pleistocene aquifer on either side of clay lenses. This points to the delaying ef-

Fig. 9. Depth to fresh–brackish groundwater interface (isopleth for 150 mg chloride per litre; source: Dufour, 2000), and maximum eastward extension of Early Pleistocene and Holocene coastlines.

fect of low-permeable layers on the free-convection salinization process by density currents.

Different conditions prevailed north of the barrier dike, where deeply incised and shifting tidal inlets caused a complete salinization in the tidal-flat area of the Waddenzee behind the barrier islands.

Effects of land reclamation

The coastal lowlands used to form extensive peat marshes behind the coastal barrier with its dunes, and locally consisted of tidal flats and salt marshes along the tidal inlets. This situation prevailed when the first land-conversion works began at about 800 AD. The peat bogs then formed large dome-shaped areas, several metres above sea level, which were dissected by distributary branches of the main delta streams. Drainage of these areas for agricultural use caused a lowering of the ground surface by compaction and disintegration of the peat. Eventually this brought the area below sea level which made it necessary to build dikes, dams and other hydraulic structures, so that a patch-work of polders with artificially controlled water levels came into being (e.g. Van de Ven, 1993).

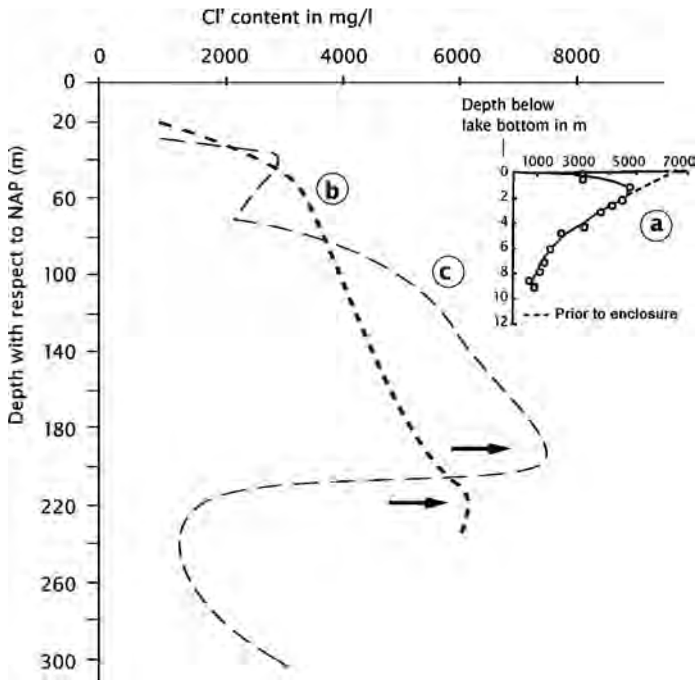

Fig. 10. Vertical chloride distribution in the area of the former Zuiderzee: (a) Holocene clay, central IJsselmeer; (b) Pleistocene sand, central IJsselmeer (after Volker, 1961); (c) Borehole 25G-32, situated east of Amsterdam at the edge of the Zuiderzee area, showing freshening of groundwater in the upper part of the marine Maassluis and Oosterhout formations below ca. 190 m. Arrows indicate clay layer at the boundary between fluvial Pleistocene and underlying marine Pleistocene and Pliocene deposits.

The reclamation of land from lakes and sea embayments was made possible by the introduction of drainage by means of windmills in the early 15th century. Large areas were regained, which previously were lost by sea encroachment, river flooding and destructive digging of peat for fuel. The lowest polders are currently more than 6 m below mean sea level. The ongoing change in topography through land and water-management measures has set in motion a complicated pattern of groundwater recharge and discharge systems. Because of the artificially maintained groundwater tables in the low polders, a continuous flow of groundwater from the relatively higher regions into the lower ones occurs, causing upward seepage ('kwel') of groundwater from the Pleistocene aquifer into the polder area (Fig. 11).

The loss of water from the Pleistocene aquifer by this upward leakage is partly compensated for by subsurface inflow of fresh water from the higher grounds in the east (causing desiccation problems at the fringe of these higher areas) and by intrusion of sea water in the west (causing salinization in the lower areas). Saline groundwater brings through upward leakage a large amount of salt to the surface, which partly precipitates in the soil by evaporation during the summer. For the area between Amsterdam and The Hague, this input of chloride into soil and water may be in the order of 150 000 tonnes annually (ICW, 1976). Flushing of the soil by the precipitation surplus in winter, and flushing of the ditches and canals by river water, protect the area against complete salinization. The process of redistribution of salt and fresh groundwater takes thousands of years to reach equilibrium with new hydraulic conditions. With the acceleration in changing conditions through human interference during the last 1000 years, it is evident that the fresh–salt water interfaces are far from steady-state. Prediction of environmental problems within the framework of this long-term hydrogeological evolution is one of the priorities in groundwater research in this area.

A schematic reconstruction of the regional redistribution of fresh and salt water due to the water-management and land-reclamation works is depicted in Figure 12 (De Vries, 1981). This section is illustrative for the area between Amsterdam and The Hague, which is dominated by a central zone with deep polders, bounded by the coastal dune belt in the west and the ice-pushed ridge of the Utrechtse Heuvelrug in the east. Fresh water from these higher grounds and saline groundwater from greater depths and from the sea, discharge into the polders through upward leakage. Local groundwater flow systems, connected with the polder topography, are superimposed on these regional systems. Groundwater flow velocity calculations indicate that the landward intrusion of the sea water by lowering of the land surface during the last 1000 years, is not more than about 6 km. Thus, most of the saline groundwater originates from earlier transgressions and from old marine deposits at greater depths.

The seepage flux is mainly controlled by the hydraulic resistance of the Holocene confining layer (parameter *c* in Fig. 5), and can reach an areal average of more than 15 mm/day along the eastern margins of the area, where the remnants of Holocene peat on the Pleistocene sand form only a thin layer and where the hydraulic gradient is high. In fact, one polder, the Béthune polder (at the western fringe of the Utrechtse Heuvelrug), produces annually 30 million m^3 of fresh leakage water, providing 30% of the total drinking-water consumption of Amsterdam. The remaining 70% is extracted from the coastal dunes where the natural replenishment is supplemented by artificial recharge with water from the river Rhine.

The Holocene confining layers consist of alternating sandy channel deposits, clayey flood-plain and tidal-flat sediments and organic swamp deposits. Thorough analyses of the variation in hydraulic properties (notably the specific vertical flow resistance) in connection with the sedimentological architecture of the fluvial Rhine-Meuse Delta, and its geostatistical characteristics, were given by Bierkens (1994) and Weerts (1996).

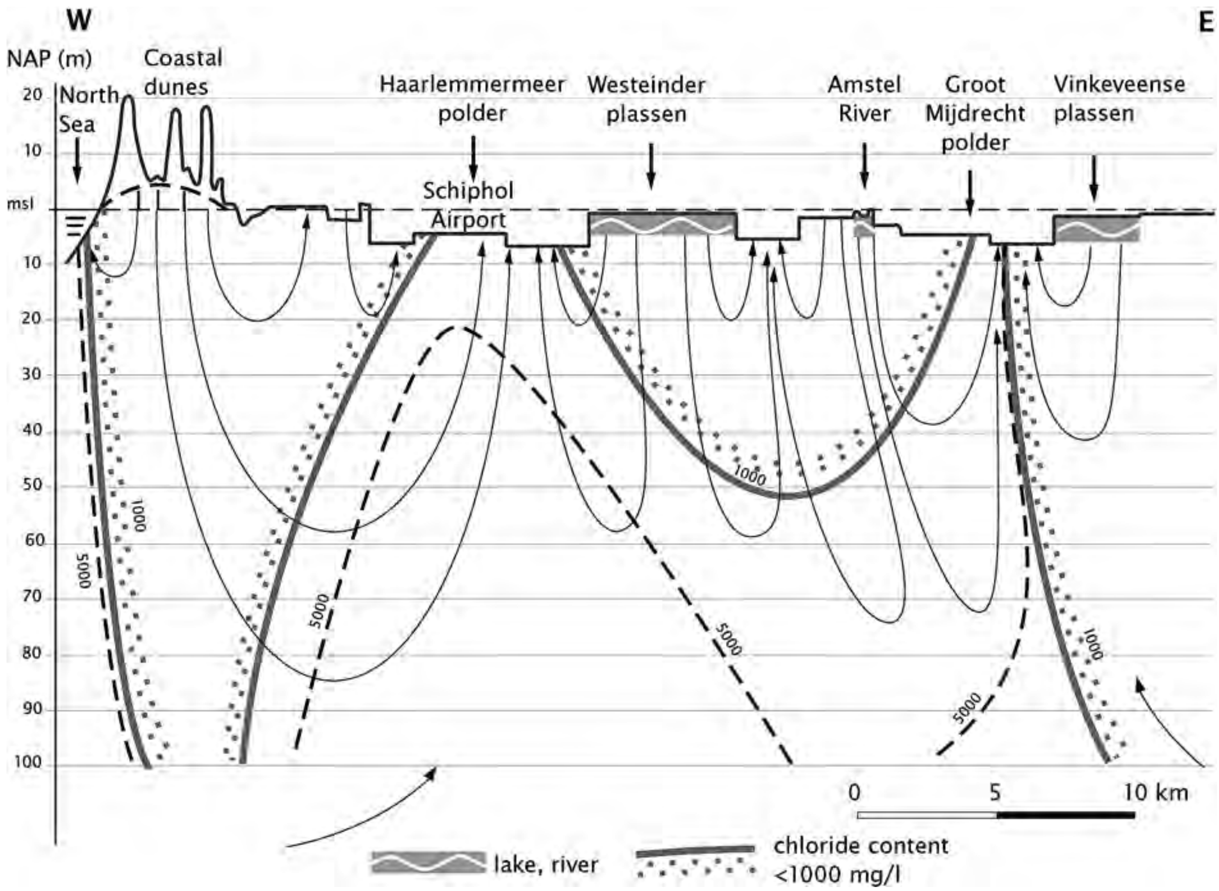

Fig. 11. Groundwater flow systems in an east-west section about 20 km south of Amsterdam, showing the strong influence of topography in this polder area (inferred from boreholes and numerical model simulation).

Process reconstruction and prediction

From the early 20th century onwards, Dutch geohydrologists have successfully developed mathematical models to describe and explain groundwater-flow conditions in leaky aquifers, for gravity as well as density-driven flow into polders and excavations, and for extraction wells (see section 'Development of groundwater research'). Equally important in reconstructing the paleohydrological evolution is the investigation of the regional distribution of hydrochemical water facies in combination with environmental isotopes. For example, freshening of salt water (NaCl water) and salinization of fresh water ($Ca(HCO_3)_2$ water) often leads to the following cation-exchange reactions at the clay or organic-material adsorption complex (Adc):

$$\text{Na Adc} + Ca(HCO_3)_2 \text{ water} \rightarrow \text{Ca Adc} + NaHCO_3 \text{ water},$$

$$\text{Ca Adc} + \text{NaCl water} \rightarrow \text{Na Adc} + CaCl_2 \text{ water}.$$

The occurrence of $NaHCO_3$ or $CaCl_2$ in groundwater thus testifies to the freshening and salinization processes, respectively. This occurs for example in the Frisian coastal area, where flushing by fresh water from the higher grounds produced $NaHCO_3$, whereas recent salt-water encroachment through groundwater extraction at the Noordbergum pumping station resulted in the occurrence of $CaCl_2$ (Fig. 13). The first to explain this origin of $NaHCO_3$ was the hydrologist and mining engineer Jan Versluys (1916). On the basis of this concept, Geirnaert (1973) prepared an overview of hydrochemical groundwater types in the west of the Netherlands in relation to the evolution of the distribution of fresh and salt water.

Appelo et al. (1990) recognized cation-exchange reactions to follow a chromatographic pattern along flow lines, and made use of this theoretical model to simulate paleohydrological processes. Laboratory experiments and isotope tracers were applied to calibrate the models. In this way, Beekman (1991) refined Volker's diffusion calculations (see section 'Hydrogeological evolution') and made a reconstruction of the diffusion and dispersion processes below the Zuiderzee bottom in relation to various phases of erosion and sedimentation since the first encroachment of the sea in medieval times. He further analysed the evolution of a 100-m-deep fresh groundwater occurrence in West-Friesland, west of the Zuiderzee area, and concluded from chromatographic analysis, chemical balances

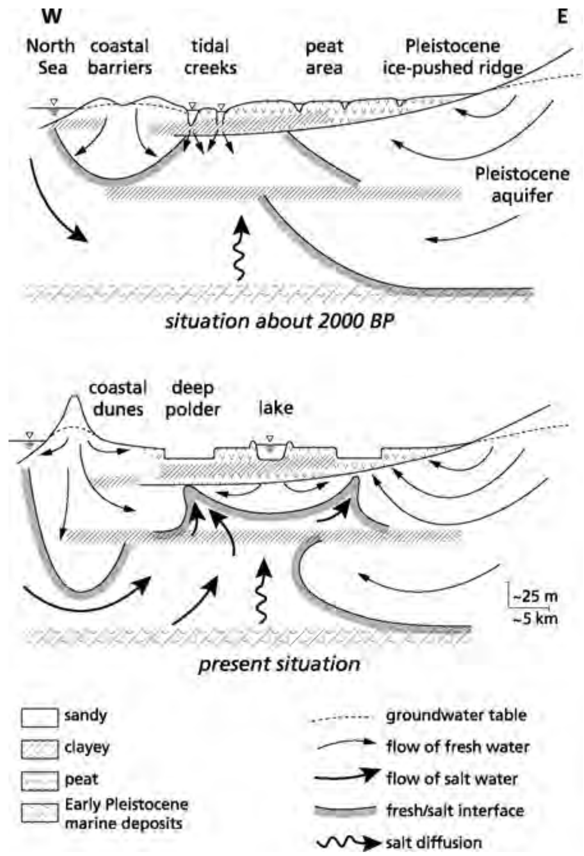

Fig. 12. Schematic 'genetic' topographic-hydrogeological sections through the area between Amsterdam and The Hague, showing the influence of land-reclamation works during the last 1000 years (after De Vries, 1981).

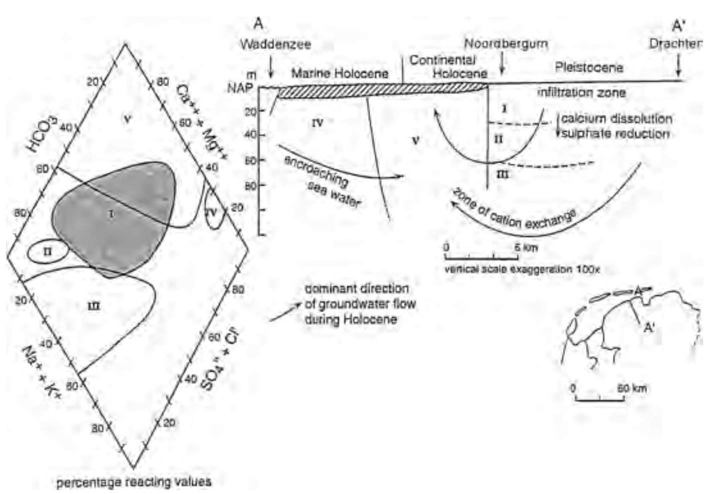

Fig. 13. Hydrochemical section across the Frisian coast and corresponding classification diagram, showing: (i) infiltration in low-carbonate soil (zone I), followed by dissolution of carbonate (II) and subsequent cation exchange with $NaHCO_3$ generation in the zone of freshening (III) and (ii) sea-water encroachment in zone IV, causing cation exchange and $CaCl_2$ generation in zone V; sea-water encroachment is caused by pumping at Noordbergum (based on data from Geirnaert (1973)).

and isotope studies an origin by local infiltration, mainly during the last 300 years. This infiltration process was initiated by the reclamation in the 17th century of the deep polders Beemster and Schermer at the southern fringe of West-Friesland (cf. Fig. 6).

Another comprehensive hydrochemical-paleohydrological study was carried out by Stuyfzand (1993) in the coastal dune area. He reconstructed the evolution of the shape of the fresh-water lens and its chemical composition as caused by land-reclamation works behind the dunes as well as by groundwater extraction and artificial infiltration of river water. This study includes an advanced hydrochemical classification system, based on the origin of a water mass (hydrosome) and the characteristic hydrochemical zones (facies) within each hydrosome. A detailed chemical and transport modelling of the evolution of water artificially infiltrated from the river Rhine into the dunes, was carried out by Van Breukelen et al. (1998). This study was based on repeated sampling for chemical and isotope analysis in more than 100 mini-tubes along a distance of 1000 m over a period of 24 years. Dominant processes proved to be: (i) seasonally dependent cation-exchange, caused by the seasonally fluctuating NaCl content of Rhine water, and (ii) net dissolution of calcite from the dune sediments due to under-saturation of Rhine water and the production of CO_2 by dissimilation processes.

The application of isotopes may be illustrated by the identification of the provenance of fresh, upward seeping water in the previously mentioned Béthune polder, adjacent to a lake (Loosdrechtse Plassen) west of the Utrechtse Heuvelrug (Fig. 7a). Groundwater in the ridge of hills reflects the average Dutch rainwater ^{18}O content of –7.5‰ V-SMOW, whereas the lake water shows a value of –4‰, due to ^{18}O enrichment by evaporation. The seepage water has an intermediate content of –6‰, indicating a mixed origin (unpublished research Vrije Universiteit Amsterdam). Stuyfzand (1993) and Meinardi (1994) used various annual peaks in post-1950 nuclear-bomb-test tritium contents in rainwater as markers in vertical groundwater profiles to identify years of infiltration and to calculate the percolation flux. An example of a paleo-hydrogeological study by a combination of hydrochemical transport modelling and isotope analysis on a time scale of tens of thousands of years is the reconstruction of conditions and processes connected with rainwater infiltration and transport in the Tertiary Ledo-Paniselian coastal aquifer in Flanders south of the Dutch border (Van der Kemp et al., 2000).

Such reconstructions are a prerequisite for explaining the present situation and establishing the initial conditions to predict future developments of groundwater flow systems and redistributions of fresh and salt water by mathematical flow-simulation modelling. Oude Essink (1996) developed a comprehensive numerical computer model to simulate future shifts in the fresh–salt water interface in the coastal area, including scenarios for sea-level rise and continuing land subsidence. He predicted that, even in the case of the absence of future relative sea-level rise, it will require at least another 5000 years to reach equilibrium in the distribution of fresh and salt water.

The Pleistocene inland area

Aquifer character

In contrast to the Holocene coastal area, the Pleistocene area has remained relatively unchanged from a geological viewpoint during the last 10 000 years. The more humid conditions during the Holocene in comparison to the Weichselian, caused expansion of the stream system on sloping areas and the growth of extensive raised bogs on poorly drained flat areas. Subsequent changes were caused by human interference, including excavation of peat bogs, reclamation of dry land from marshy areas, canalization of rivers, regulation of runoff, and extraction of groundwater for public water supply.

The Plio-Pleistocene aquifer is almost everywhere filled with fresh water. Brackish groundwater is restricted to the marine, Early Pleistocene and Tertiary deposits (Figs 7, 9). The physico-chemical character of fresh groundwater normally changes from light-acidic to light-alkaline on its way from the recharge to the discharge area. Infiltrating rain water has a Na,K-sulphate/chloride character. Production of carbon dioxide and loss of oxygen by dissimilation in the upper soil layers subsequently cause the groundwater to dissolve calcareous material into bicarbonate and to reduce sulphate to sulphide. Most groundwater in the Pleistocene aquifer is therefore dominated by a Ca,Mg-bicarbonate water type. Mobilization of iron under conditions of low pH and/or low redox potential occurs particularly in marshy areas. Groundwater exfiltration in brooks in such areas has locally produced bog iron-ore deposits.

The exposed phreatic aquifers in the Pleistocene area are susceptible to contamination. Shallow groundwater is polluted diffusely by agricultural activities and atmospheric contaminants, and locally by industries and waste dumps. Nitrate concentrations often exceed the drinking-water standard of 50 ppm; concentrations up to 400 ppm have been observed at some locations as a result of manure deposition from high-intensity pig breeding. Denitrification of nitrate delivers oxygen, which in contact with pyrite (FeS_2), that is often abundant in marine sediments, can produce sulphuric acid according to the reaction:

$$2\,FeS_2 + 7\,H_2O + 15\,O \rightarrow 2\,Fe(OH)_3 + 4\,H_2SO_4.$$

This results in a low pH, which in turn can mobilize toxic metal ions.

Tritium data from the national groundwater-quality network have been used to investigate the relation between age stratification and pollution. This analysis revealed that the proportion of young (post-1950) groundwater in the upper 15 m of the sandy aquifer strongly decreases from more than 75% in the recharge areas to less than 25% in the discharge areas. The dense superficial drainage system in the low-lying discharge areas has removed a substantial part of recent water but prevents replenishment of the deeper aquifers (Broers, 2002).

Groundwater abstraction from deep aquifers causes salinization and termination of groundwater seepage in valleys with valuable groundwater-dependent ecosystems. Shallow nutrient-rich groundwater thus displaces pristine groundwater, and rare oligothropic ecosystems are accordingly replaced by ordinary euthrophic communities (Stuurman, 2000). For an overview of anthropogenic influences on groundwater quality and the associated geochemical and microbial processes, reference may be made to Griffioen et al. (2003).

Local and regional groundwater and stream systems in level areas

The Pleistocene deposits outside the ice-pushed hills form a slightly undulating north- and westward sloping topography with an overall 1 : 2500 gradient (Fig. 1). Fluvial coarse-grained strata are covered by a layer of peri-glacial and fluvio-eolian, fine-grained sandy and loamy deposits with a maximum thickness of 10 m. Groundwater in the undulating areas is normally part of a hierarchy of nested groundwater flow systems. It is characterized by deep regional systems between the higher-order topographic elements, and by shallow, local or intermediate systems, which are driven by lower-order topographic elements like brook valleys and small streams and ditches (Engelen & Kloosterman, 1996).

The area is drained by a hierarchical stream system that expands and contracts with the seasonal fluctuation of the groundwater table (Fig. 14). This fluctuation thus regulates the number of channels that participate in the drainage process, so that the drainage density increases with increasing precipitation surplus. The maximum groundwater discharge capacity of a drainage system ranges between the average annual rainfall surplus of about 1 mm/day for the relatively higher areas with the deepest groundwater tables (> 5 m below surface), to 12 mm/day for the relatively low and flat areas with shallow groundwater tables (< 0.5 m below surface). This connection between decreasing groundwater depth and increasing discharge rate is due to the reduction in storage capacity with decreasing groundwater depth as well as to a reduction of the hydraulic-head gradient under low and flat topographic conditions.

De Vries (1974, 1994) characterized this topography-

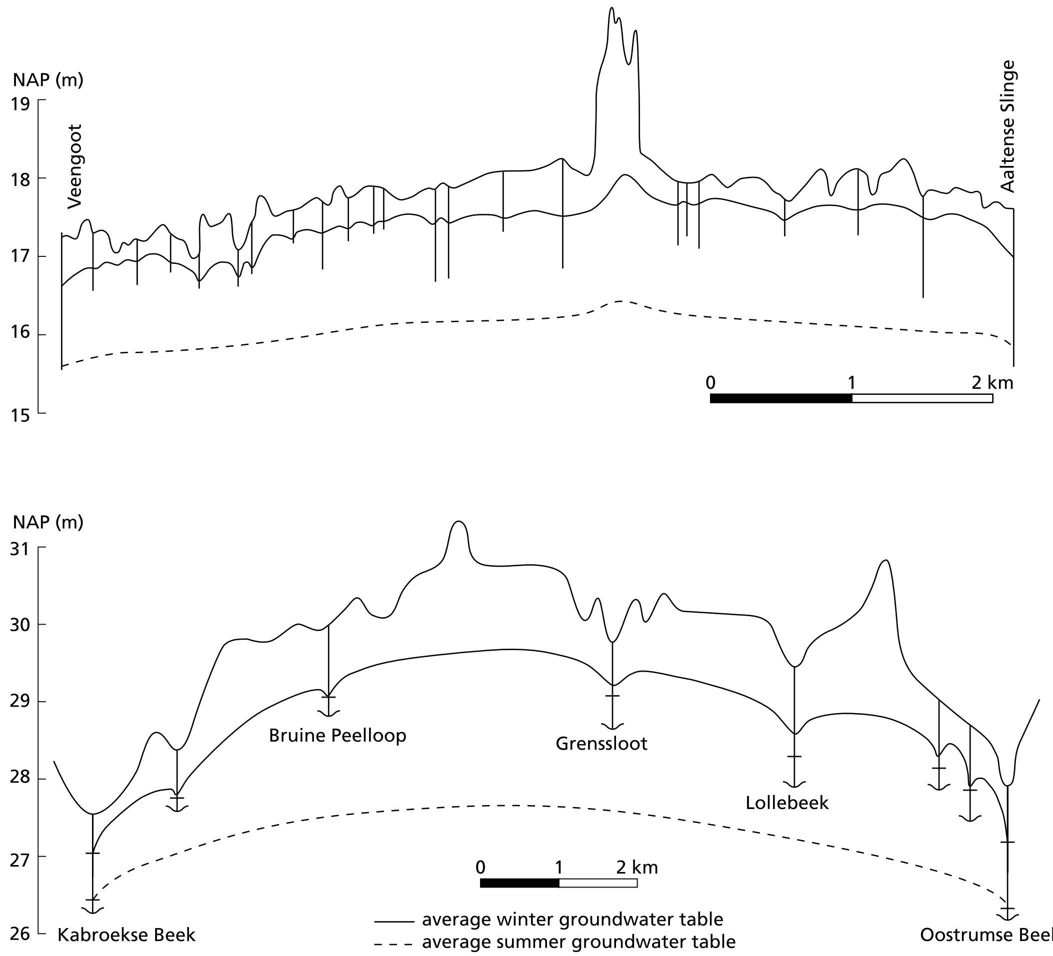

Fig. 14. Hydrological-topographic sections through two small stream systems in the Netherlands Pleistocene area, eastern Gelderland (above) and northern Limburg (below), indicating: (i) higher stream density with decreasing groundwater depth, and (ii) seasonal expansion and contraction of the discharge-contributing stream system, regulated by the fluctuating water-table depth (after De Vries, 1994).

controlled drainage system as an expression of the continuum of groundwater and surface water in which streams are initiated by groundwater seepage and sapping erosion at the intersection of groundwater table and topography in such way that the stream spacing corresponds to the required discharge capacity.

An interesting phenomenon in the area of the Roer Valley Graben is the influence of faults on groundwater flow in the Pleistocene sandy aquifer. Smearing and mineral deposition on the fault planes have created high resistance to horizontal flow. The result is a relatively steep hydraulic gradient across the fault with shallow, obstructed groundwater on the elevated block at the upstream side of the groundwater flow system. These unusual, wet soils on higher grounds are locally known as 'wijstgronden' (Bense, 2004). Exfiltration of groundwater upstream of the fault and associated sapping erosion have locally influenced the stream network.

Regional groundwater systems in higher areas

Groundwater in the more elevated areas, like the ice-pushed ridges, constitutes regional groundwater recharge systems with deeper groundwater tables. These areas are characterized by the absence of a shallow, lower-order drainage system. Such regional systems are only moderately susceptible to rainfall events and even seasonal effects, because the storage capacity is large enough to accommodate seasonal precipitation-surplus differences. Seasonal groundwater-level fluctuations are thus moderate because of the buffering effect of the thick soil–water retention zone. Main fluctuations are caused by the long-term imbalance between recharge and discharge in years of exceptionally high or low precipitation as well as by long-term climatic fluctuations (Fig. 15). Time lags between recharge and reaction of the groundwater table are of the order of one year because of (i) the time it takes the soil–water pressure wave to propagate through the soil–water zone (about 1 month per 10 m), and (ii) the delay

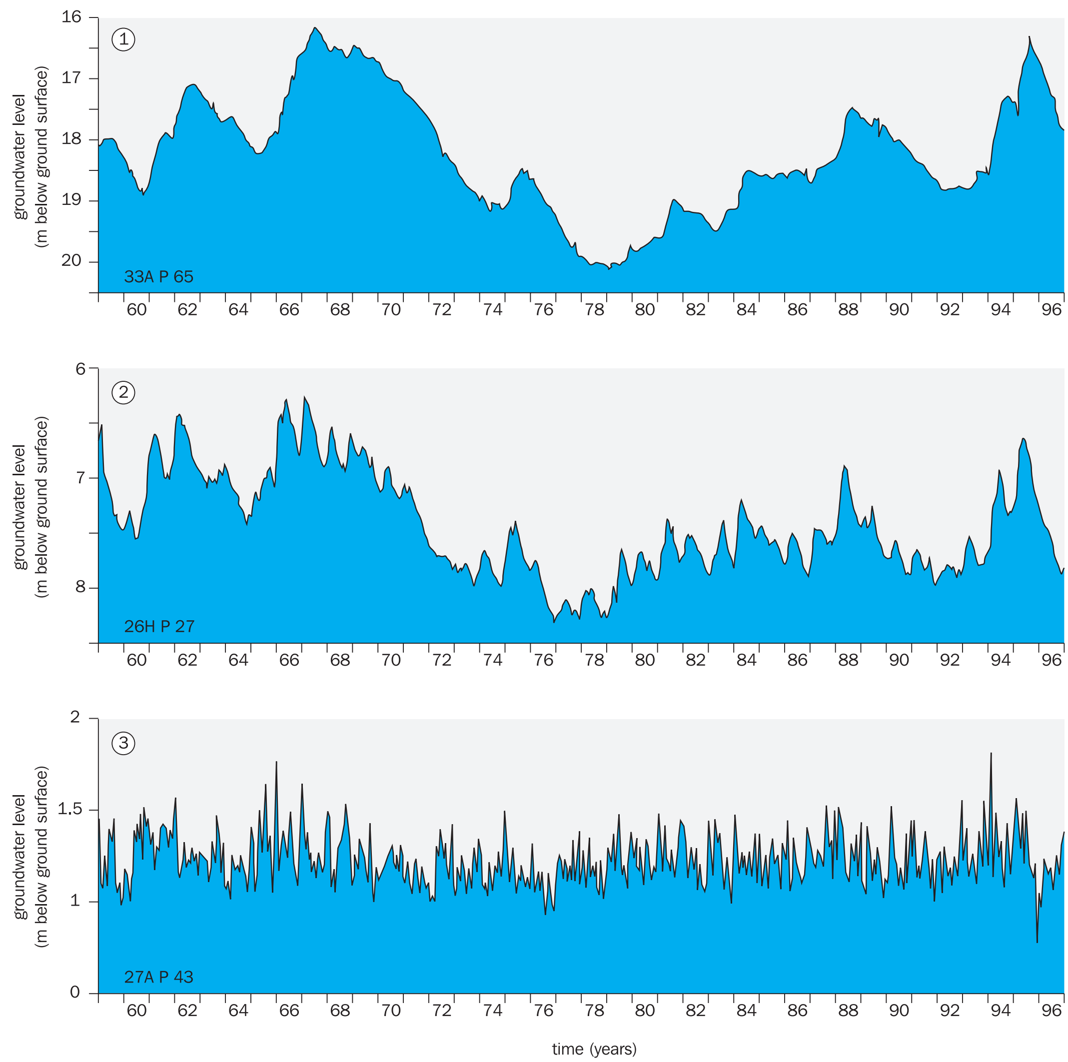

Fig. 15. Characteristics of groundwater-level time-series for three localities in the Veluwe ice-pushed ridge. This figure illustrates the buffering effect of groundwater depth (thickness of the unsaturated zone) on the impact of short-term, seasonal and annual rainfall fluctuations on the groundwater level. The absence of annual fluctuations in the shallow aquifer is due to the more or less artificially maintained regional groundwater table in this low-lying area as well as to the continuous subsurface inflow at the fringe of the ridge (after Van Bracht, 2001).

in the groundwater-level oscillation itself, due to the high drainage resistance of these large groundwater systems (Gehrels, 1999; De Vries, 2000). Conditions are different on the chalk plateau of Zuid-Limburg. The karstic chalk is characterized by a spatially irregular, but often fast response of the groundwater table to rainfall, and by strong groundwater-level fluctuations of sometimes more than 10 m (Jongmans & Van Rummelen, 1935).

Gehrels et al. (1994) carried out a time-series analysis of groundwater-level fluctuations in the Veluwe and could distinguish the impact of the reclamation in 1956 of the polder Oostelijk Flevoland (540 km^2) in the adjacent IJsselmeer from climate-induced fluctuations. Gehrels determined that it took about 25 years for the reclamation-induced head decline to arrive at the centre of the Veluwe, over a distance of 25 km. An interesting aspect of the groundwater recharge process in the Veluwe area is that the overall ^{18}O content of the groundwater (–7.8‰ V-SMOW) closely reflects the average annual content of local rainwater, although the rainfall surplus of about 350 mm originates mainly in the winter season. This winter rainfall has an average ^{18}O content of precipitation that is about 1‰ below the annual average. The explanation is a strong dispersion of seasonal rainwater in the root zone of shrubs and trees (Gehrels, 1999).

Another large-scale and supra-regional groundwater system occurs in the deep confined aquifer systems of the Roer Valley Graben (see section 'Geological frame-

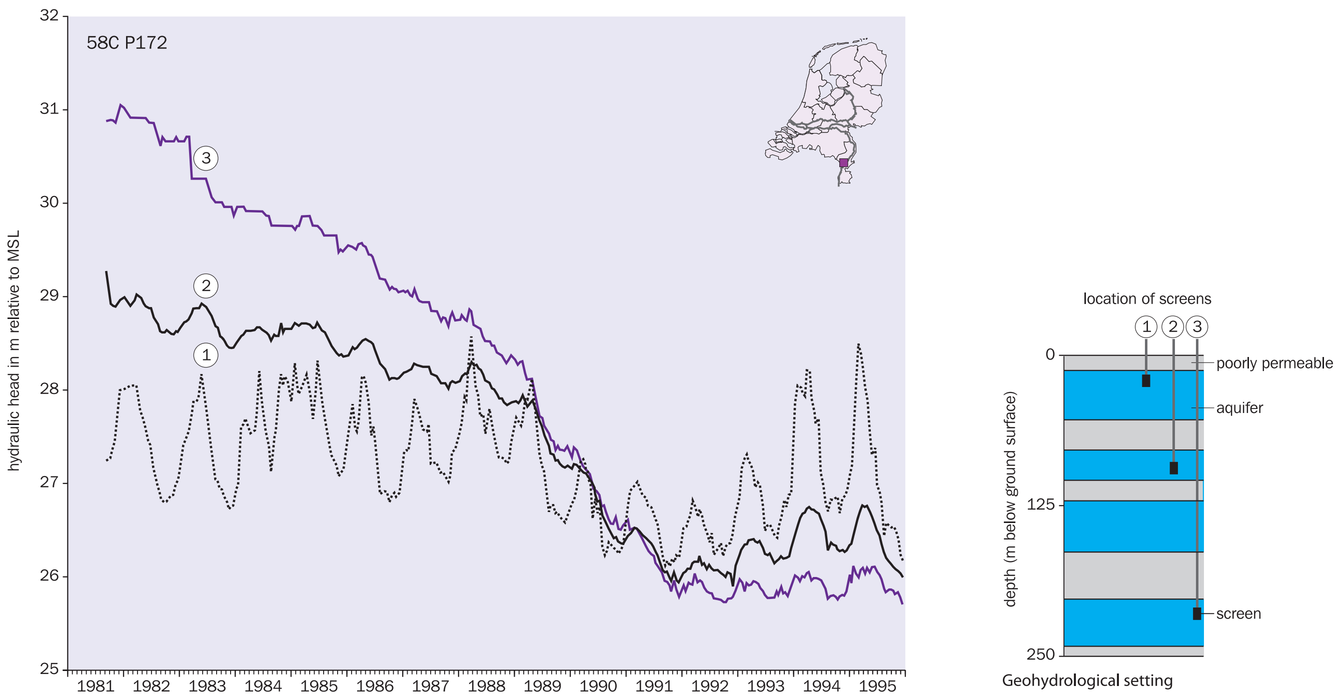

Fig. 16. Hydrographs for three screens in observation well 58C-P172 in the Roer Valley Graben in central Limburg. Groundwater extraction from the deepest aquifer in nearby Germany, caused the flow direction to change from upward seepage to downward infiltration; MSL is mean sea level (after Van Bracht, 2001).

work'). Figure 16 depicts the influence of groundwater extractions in the German lignite mines on the hydraulic head of the deep, confined Tertiary aquifers in the Dutch section of the graben. Accumulated extractions are in the order of 100 million m³ of water per lignite excavation. This cross-border decline in hydraulic head has only propagated through the confined elastic aquifers; the upper aquifers are unaffected. This caused the groundwater system to change from a condition with upward seepage into a situation with downward infiltration. The loss of seepage of old deep water with its specific chemical composition and high pH, has locally affected the natural vegetation in brook valleys in this area (Stuurman, 2000; Griffioen et al., 2003).

Relicts of paleo-conditions were encountered in the Early Pleistocene sediments beneath the ice-pushed ridges and their surrounding areas. For instance, the groundwater chemistry in borehole 32G-137 in the Eemvallei, 10 km SE of Amersfoort, shows the influence of a freshening in the upper part of the marine Maassluis Formation, with a downward increase of chloride to more than 6000 mg/l at a depth of 285 m. The ^{18}O content of this brackish water is between –8.5 and –9.5‰ (corrected for the seawater component), suggesting cold conditions during infiltration. Since ^{14}C age indications of this water are in the order of 10 000 to 20 000 years, it is plausible that the formation was partly flushed during the Weichselian by deep regional subsurface inflow from the adjacent ice-pushed hills (Meinardi, 1975). Borehole 25G-132 at the northern fringe of these hills (Fig. 10c) probably shows fresh-water relicts of the same process.

Large infiltration areas like the Veluwe and the Utrechtse Heuvelrug show relatively low subsurface temperatures, due to deep penetration of relatively cold rain water. The subsurface temperature in the Veluwe is less than 10°C at a depth of 125 m, whereas the average temperature in the surrounding areas is generally in the order of 13°C. Relatively high temperature anomalies are also observed in connection with groundwater ascending from greater depths along fractures, and in relation with rocks of high heat-conductivity, notably updoming evaporites in the northern Netherlands (Dufour, 1998).

Groundwater use and legislation

Groundwater exploitation

The total annual volume of groundwater extracted by water-supply companies for public and industrial use is in the order of 1000 million m³, of which about 18% originates from artificially induced recharge of river water, mainly into the dune area. Industry itself produces ca. 300 million m³, including water for cooling purposes (excluding cooling for power plants). Agriculture extracts 300 to 400 million m^3 from private wells for irrigation during dry summers. Most groundwater for the drinking-water supply is extracted from the Pleistocene area; it equals about 15% of the total precipitation surplus that this area receives. Considering that only part of this surplus recharges the deep aquifers and taking into account other restrictions, it is estimated that the potentially extractable resource is about twice the present extraction. Thus there is no overall over-exploitation, but nevertheless groundwater extraction has locally caused a drawdown of the groundwater table. This local drawdown together with the regional influence of intensified land drainage and reduced infiltration in built-up areas, has resulted in an overall lowering of the groundwater table in the Pleistocene area in the order of decimetres over the last 50 years (Dufour, 1998). Damage to valuable wetlands was one of the effects.

Groundwater production in the Holocene area is mainly restricted to the flood plains of the larger rivers and to induced river infiltration along stream beds. In addition, artificially infiltrated river water in the coastal dunes is an important source of drinking water for the cities in the coastal region. A special problem in this region is the necessity to maintain a shallow water table in urban areas, to protect the often centuries-old wooden foundation piles from decay.

Legislation

No special regulations with respect to groundwater had been developed until the mid- 20th century. Before that time every landowner could sink a well and extract water as long as no damage was done to the property of others. Limitations for the exploitation of groundwater were thus set by private law. This posed a problem to water supply- companies who could never be sure of continuity. Therefore the Grondwaterwet Waterleidingbedrijven (Groundwater Act Water Supply Companies) was enacted in 1954. From then on drinking-water companies needed a licence for any abstraction, specifying conditions and damage compensation; groundwater abstractions by others, however, were not authorized under this law. In 1981 an overall Grondwater Wet (Groundwater Act) became effective for all abstractions and activities related to infiltration and groundwater recharge. According to this law, the provinces are the authorities responsible for permission, registration and reporting. Quality aspects are mainly related to protection of recharge areas. Other groundwater-quality issues are dealt with in the Wet Bodembescherming (Soil Protection Act) of 1987, which includes regulations for prevention of subsurface pollution and for remediation of contaminated soils. Within this legislation, the provinces are obliged to set up and maintain a groundwater monitoring and management plan (Dufour, 1998, 2000).

Appendix

Some theoretical groundwater flow concepts (Fig. 17)

Quantitative groundwater hydrology was founded by Henry Darcy in 1856 with the results of his experiments on groundwater percolation through a vertical cylinder filled with sand. Figure 17 depicts this experiment with free outflow under atmospheric pressure. Darcy concluded that the flow rate per square metre of surface perpendicular to the flow path was proportional to the loss of hydraulic head Δh, inversely proportional to the distance e between the upper and lower flow boundary, and proportional to a coefficient k, depending on the nature of the sand. This principle that holds for all porous media is known as *Darcy's Law*; in formula:

$$q = k\Delta h/e,$$

with $\Delta h = h_1 - h_2$, where q is flow rate ($m^3/day/m^2$); h_1 and h_2 are *hydraulic head* at the upper and lower boundary of the sand layer respectively (m); k is *permeability factor* or *hydraulic conductivity* (m/day) in vertical direction; e is length of flow-line path (m). Hydraulic head is elevation head plus pressure head; assuming atmospheric pressure (zero pressure) at the lower boundary and the elevation at this boundary as arbitrary zero reference level, then the hy-

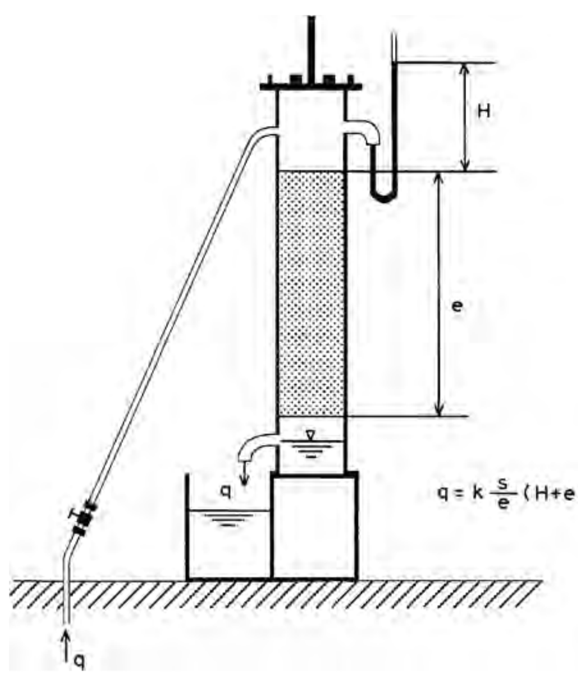

Fig. 17. Principle of percolation experiments according to Harting (1852) and Darcy (1856); see Appendix for explanation. s = surface area of cylinder.

draulic head $h_1 = H + e$ and hydraulic head $h_2 = 0$, where H is the hydraulic pressure at the upper boundary, thus:

$$q = k(H + e)/e.$$

For horizontal flow through an aquifer, the total flow rate per metre width is proportional to the product of the permeability factor k in horizontal direction and the thickness D of the aquifer. This product kD (m^2/day) is termed *transmissivity*, and often indicated with symbol T (see also Fig. 5). For vertical flow (leakage) through a semi-confining layer, the flow rate per m^2 is proportional to the hydraulic head difference between the upper and lower boundary of the confining layer, proportional to k' in vertical direction, and inversely proportional to the thickness D' of the confining layer, thus:

$$q = \frac{k' \Delta h}{D'}.$$

D'/k' is often termed *vertical flow resistance c* (day) of the confining layer (see also Fig. 5). Pieter Harting (1852) carried out similar experiments, but wrongly left out the elevation-head term e in h_1 (see 'Development of groundwater research in the Netherlands').

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Surface mineral resources

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ABSTRACT

Dutch unconsolidated Quaternary and Tertiary deposits constitute major resources of sand, gravel and clay, exploited mainly for construction works and the building-materials industry. Limited surficial occurrences of Cretaceous chalk and Triassic limestone and dolomite are carbonate resources, used mainly for the production of cement. From 1999 to 2003, an average amount of 90 Mt of sand, 4.9 Mt of gravel, 3.9 Mt of clay and 1.6 Mt of carbonate rock was extracted annually. Between 10 and 14 Mt/a of the aggregates production was exported; imports of aggregates rose from 30 to 45 Mt/a. For reasons of sustainability, policy has been developed to stimulate the application of secondary building and construction materials, i.e. recyclable waste materials and industrial by-products, as alternatives for natural materials. Their total use rose from 7 Mt/a in the early 1980s to ~32 Mt/a in the early 2000s.

Keywords: Netherlands, aggregates, gravel, sand, clay, limestone, shells, secondary materials, mineral extraction, mineral planning

Introduction

Dutch industry currently requires about 159 Mt/a of aggregates (sand, gravel, crushed rock etc.), clay and carbonates, mainly for construction purposes and as raw building materials (Fig. 1). About 101 Mt/a is quarried in the country and around 13 Mt/a of this production is exported. 71 Mt of the total demand is met by imports of primary (natural) aggregates and by home production of secondary materials, i.e. recyclable industrial by-products and waste materials.

The Dutch shallow subsurface consists almost entirely of unconsolidated Quaternary clastic deposits (Fig. 2). Their general physiographic setting is determined by uplifted sediment source areas in the catchments of the Rhine and Meuse rivers, south and southeast of the Netherlands, and by the subsiding North Sea Basin in the west and north. Accordingly, Quaternary sediments grade, in general, from coarse-grained in the southeast towards fine-grained in the western and northern parts of the country. The occurrence of bedrock at or near the surface, mostly Mesozoic carbonates, is limited to southern Limburg and easternmost Gelderland and Overijssel (Fig. 2). This chapter focuses on the main surface mineral resources, their extraction and related societal topics.

Commodities and resources

Concrete and mortar sand

APPLICATIONS AND DEMAND

Concrete and mortar sand ('beton- en metselzand') refers to a range of industrially produced coarse sands. About 75% is used for concrete production and about 15% in masonry mortars. The remaining ~10% is used for a wide range of purposes such as the production of asphalt and bricks, and for unconsolidated applications such as drains and filters. Each of these applications has specific requirements for the grain-size distribution and, to some extent, for the grain shape of the aggregates (Fig. 3). Producers have to sieve and blend natural aggregates in order to meet these requirements.

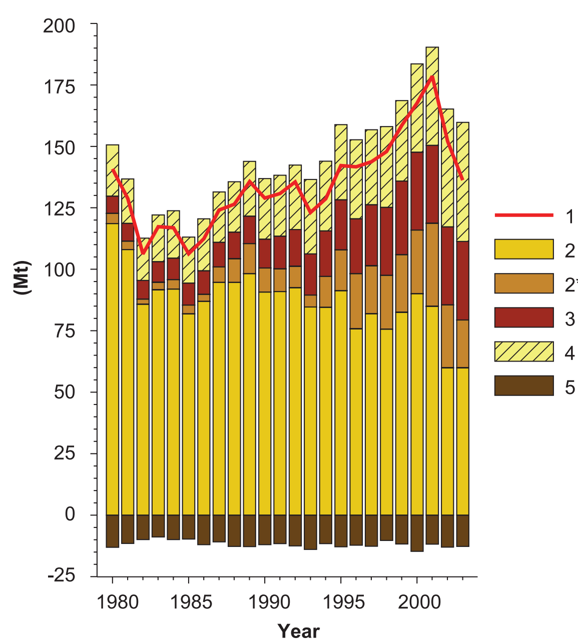

Fig. 1. The demand for aggregates, clay and carbonates in the Netherlands from 1980 to 2003 (1), and its provision by home production (2; 2* is the share of North Sea extraction), secondary materials (3) and imports (4). The growing demand was met with increasing amounts of sea-won, imported and secondary materials. Exports (5, shown as negative values) have remained more or less constant. Sources: National Commission for the Co-ordination of Mineral Planning Policy (LCCO, The Hague) and data at the disposal of the authors. Updated statistics will be published on the internet, see http://www.delfstoffenonline.nl.

Fig. 2. Geological map of the Netherlands, showing the provinces, the rivers IJssel (A), Rhine (B) and Meuse (C), the Waddenzee back-barrier basin (1), the IJsselmeer lake (2) and the Westerschelde estuary (3).

The current Dutch demand for concrete and mortar sand is about 24 Mt/a (Fig. 4). From the early 1980s till the late 1990s, the home production has been kept in approximate balance with the demand. For geological reasons, however, the share of coarse concrete sand in the total production has always been too small to meet the demand of the concrete producers. Therefore, about 7 Mt/a of coarse concrete and mortar sand were imported from Germany, while a similar amount of home-produced finer concrete and mortar sand was exported to Flanders (Belgium), which has limited coarse-sand resources altogether. From 2000 onwards, the home production decreased, and imports began to exceed exports significantly: net imports amounted to about 11 Mt in 2003.

RESOURCES

A first-order identification of natural sand resources suitable for concrete and mortar sand production, is usually accomplished by mapping sands with a median grain size of 300 μm or over. Most of this sand occurs in the southern and eastern parts of the Netherlands (Fig. 5). The bulk is of Pleistocene age and occurs in the formations of Kreftenheye, Urk, Beegden and Sterksel (all formations referred to are cf. De Mulder et al., 2003). The coarse-sand occurrences protrude westward below the Rhine and Meuse river plains, and into the North Sea (Kreftenheye and Urk formations). However, since Pleistocene river sands are overlain by westward thickening Holocene fine-grained sediments, and also display downstream fining, their exploitability in the coastal and North Sea areas is questionable (Van Heijst et al., 2004).

Coarse-sand extraction sites exist throughout the eastern and southern Netherlands, but since much of the sand is shipped to the densely populated western part of the country and to Belgium, most of it is extracted in the vicinity of the main rivers for ease of transportation (Fig. 5).

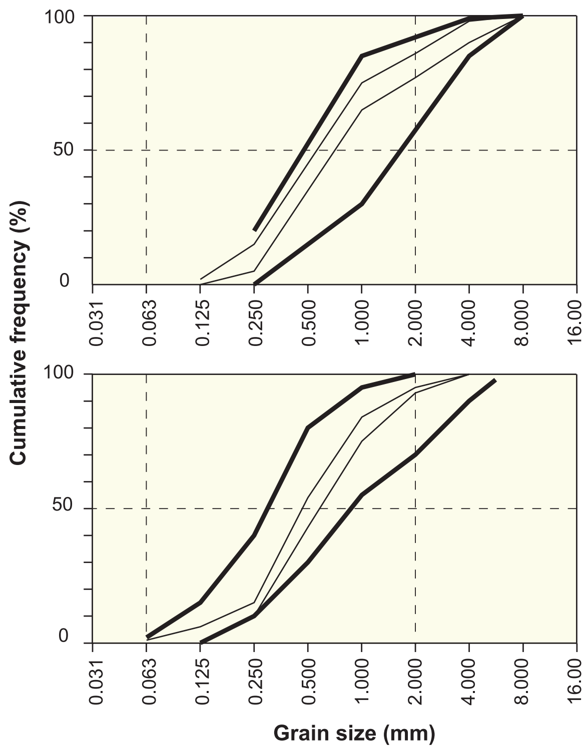

Fig. 3. Grain-size distributions of the most important types of mortar sand (lower panel) and concrete sand (upper panel). Thick lines show the official minimum and maximum requirements (NEN, 1991; 1999); thin lines delimit the actual distributions on the market (Postma et al., 1998 and unpublished data).

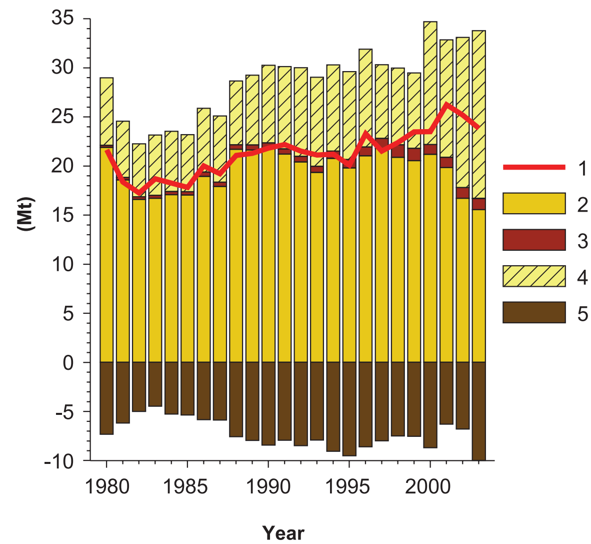

Fig. 4. The demand for concrete and mortar sand in the Netherlands from 1980 to 2003 (1), and its provision by home production (2), secondary materials (3) and imports (4). Imports (mainly from Germany) and exports (5; mainly to Belgium) are more or less in balance. Sources: see Fig. 1.

Filling sand

APPLICATIONS AND DEMAND

The western and northern parts of the Netherlands have a largely muddy or peaty subsurface. Building in these areas tends to result in compaction of the underlying soil, and therefore requires large amounts of fill as a foundation (De Jong & De Mulder, 1998). The current yearly demand for fill is about 60×10^6 m^3. The demand is met with filling sand ('ophoogzand'), of which $\sim 47 \times 10^6$ m^3 is extracted yearly (Fig. 6, panel A), and with various secondary materials.

The annual extraction of filling sand from the Dutch sector of the North Sea rose from approximately 2×10^6 m^3 around 1980 to over 18×10^6 m^3 in the early 2000s (Stolk & Seeger, 1999; Fig. 6, panel A). This increase is due to a policy of phasing out the extraction in the densely populated provinces Noord-Holland and Zuid-Holland, and more recently in Noord-Brabant. Approximately 6×10^6 m^3/a of filling sand is extracted from the IJsselmeer and a similar amount is produced as a by-product of concrete and mortar sand. Specialised extractions still remain an important source for filling sand. Such relatively small pits, which mainly supply local markets, exist all over the eastern and southern parts of the country.

RESOURCES

In contrast to concrete and mortar sand, filling sand has virtually no grain-size constraints and includes all sands which are unsuitable (i.e. too fine-grained) to be used for more demanding applications. Fine sand is available all over the Netherlands (Fig. 5). The Dutch Holocene comprises a wide range of fine-grained sand bodies, deposited in sedimentary environments ranging from fluviatile (e.g. Echteld and Beegden formations) to shallow marine (Naaldwijk and Southern Bight formations). Resources in the North Sea and the IJsselmeer are virtually unlimited.

In Limburg, hoggin ('stol') is used for about 7% of the provincial filling-material provision. It is extracted from fine-grained soils which developed on fluviatile pebbly sands of the Miocene–Pliocene Kieseloolite Formation. The pebbly sand-loam mixture is relatively cohesive, unsuitable for concrete and mortar sand production, but apt for use as filling or road-foundation material.

Sand for coastal nourishment ('suppletiezand')

The current Dutch coastal defence programme, initiated in 1990, involves monitoring the coast line and coastal processes, and measures at places where the coast is eroded. Sustained coastal erosion is countered by the application of sand on the beach or shoreface. Coastal nourishment requires between 5 and 16×10^6 m^3/a of sand

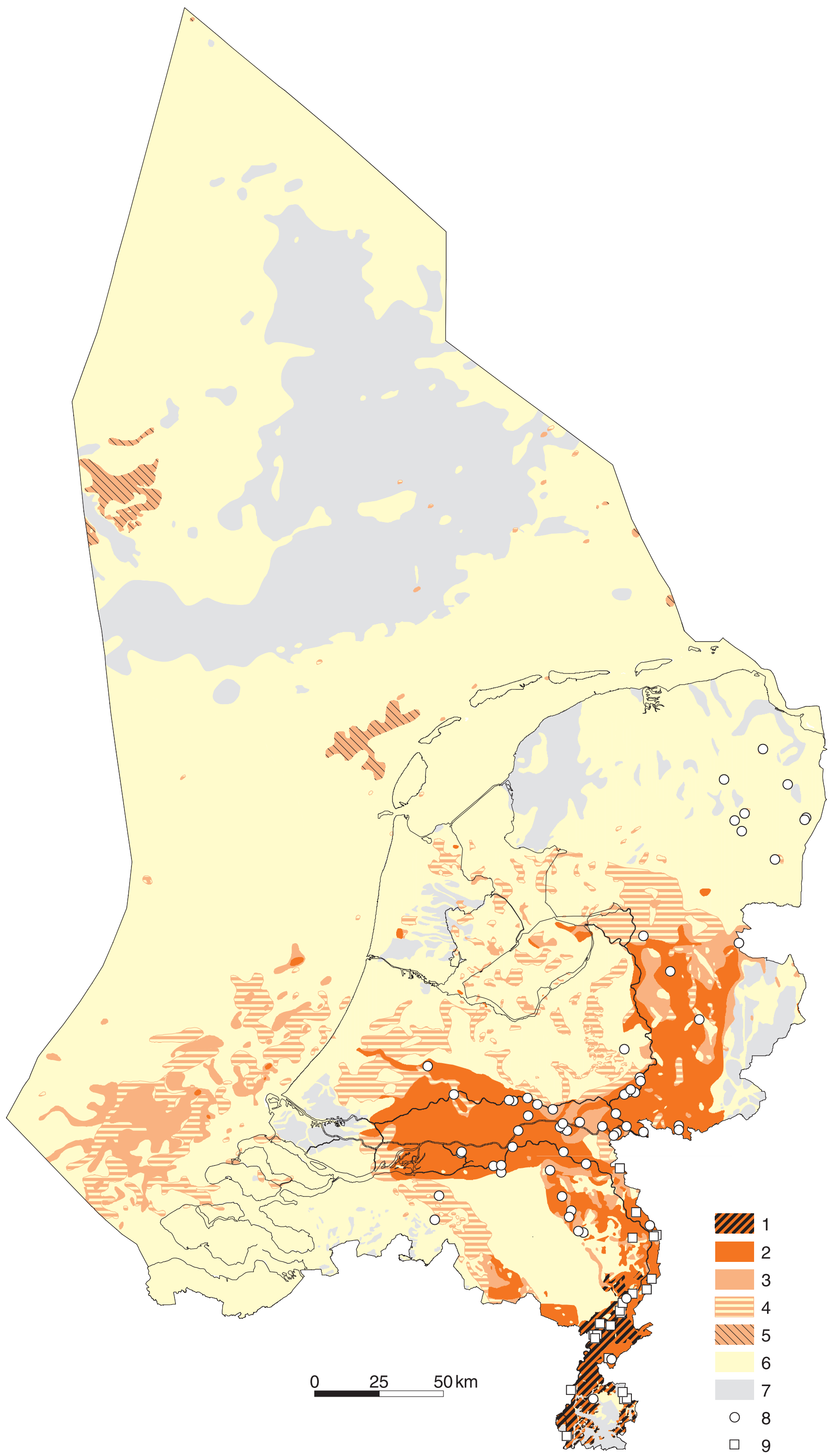

Fig. 5. Main occurrences of aggregates in the shallow Dutch subsurface (i.e. down to ~30 m below the land surface or sea bottom): (1) gravel; (2) coarse sand (cumulative thickness > 10 m); (3) coarse sand (cumulative thickness 5–10 m); (4) coarse sand and moderately coarse sand (cumulative thickness > 5 m); (5) thin surficial occurrences of coarse sand and gravel in the North Sea (thickness < 2 m); (6) fine sand; (7) no or limited aggregates; (8) extraction site of concrete and mortar sand; (9) gravel extraction site. See also: http://www.delfstoffenonline.nl.

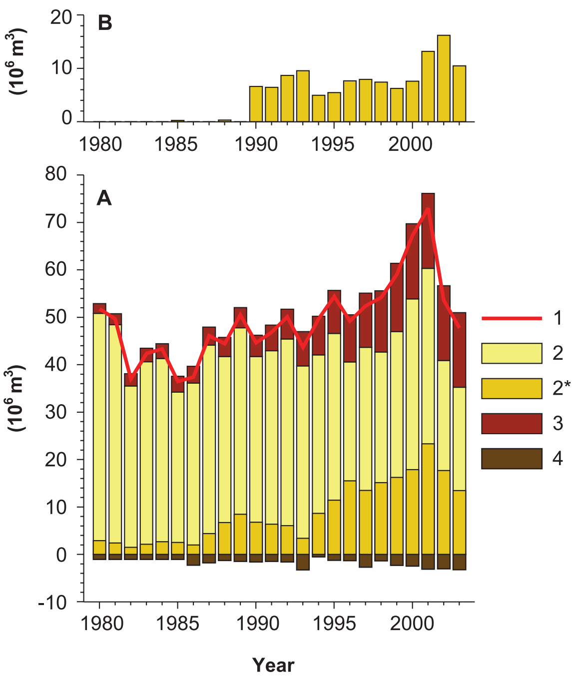

Fig. 6. Panel A: the demand for fill (1), and its provision by home production of filling sand (2; 2* is the share of sea-won filling sand) and secondary materials (3) in the Netherlands from 1980 to 2003. As a result of a restrictive policy for on-land extraction, the share of filling sand from the North Sea has grown considerably. Limited amounts of filling sand are exported (4), imports are nil. Panel B: the extraction of sand for coastal nourishment. Amounts of filling sand and sand for coastal nourishment are usually given in volume rather than in weight units. Sources: see Fig. 1.

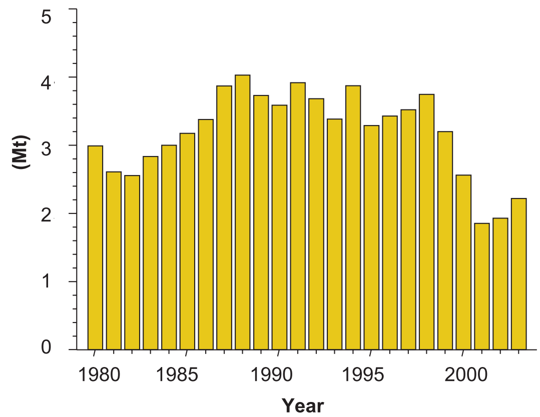

Fig. 7. Production of sand for sand-lime bricks in the Netherlands from 1980 to 2003. Import figures are not available, but are expected to be low. Sources: see Fig. 1.

(Fig. 6, panel B), which is extracted at sea as near as possible to the nourishment site.

Sand for sand-lime bricks

Sand-lime bricks ('kalkzandstenen') are produced from a mixture of sand (~90%), lime (~7%) and water, which is moulded under mechanical pressure and hardened under steam pressure and heating. They are mainly used for walls inside buildings. Sand for sand-lime bricks ('kalkzandsteenzand') has a median grain size between 150 and 400 μm. Ten to 30% in the grain-size distribution has to be finer than 125 μm and an equal share coarser than 500 μm. Natural sands which meet these requirements are quite common in the Netherlands. Most of it is extracted in the vicinity of the sand-lime brick factories, which can be found throughout the country. The production is about 2.5 Mt/a (Fig. 7).

Silica sand

APPLICATIONS AND DEMAND

Silica sand ('zilverzand') is a quartz resource for several industries. Various types of silica sand, with minimum SiO_2 percentages ranging from 97.0 to 99.8%, are used in the glass industry (~55%), the chemical industry (15%; for production of, e.g. water glass, zeolites and carborundum (SiC)), the ceramic industry (~20%) and foundries (~10%). Lower-grade silica sands are used when a white colour is desired for, e.g. masonry, the sand joints in pavements, or the production of sand-faced bricks. Table 1 shows the quality requirements for the main applications of silica sands in terms of quartz, iron, aluminium and titanium contents, and loss on ignition. Additional quality parameters, not shown in the table, include the grain-size distribution, and the chromium (Cr) and heavy-mineral contents.

Natural silica sands usually do not meet the industrial requirements and have to be upgraded. Impurities related to specific grain-size fractions such as clay, which has a high aluminium content, are removed by washing or wet screening. Grain coatings of iron minerals such as limonite can be removed by attrition (scrubbing), or acid leaching (both cold and hot). Froth flotation and magnetic or gravity separation are used to remove discrete contaminant minerals. The silica sands currently extracted in the Netherlands usually only require washing, or the combination of washing and attrition.

The production of silica sand was around 0.4 Mt/a in the 1980s and has risen to around 0.8 Mt/a in the early 2000s (Fig. 8). The rise is largely due to industrial upgrading of lesser-quality silica sands, which are relatively abundant. There is a significant export to Germany; silica sand is also imported from the large-scale extraction site at Mol (Belgium) and from several sites in Germany. Because of its high market value with respect to other Dutch quarry products (Table 2), silica sand is transported over relatively large distances. The highest-grade Dutch silica sands are transported as far as southeastern Germany and northern France.

Table 1. Matrix of five categories of silica sand (1–5) and their applications.

Categories	Applications							Chemical characterisation				
	A	B	C	D	E	F	G	SiO_2	Fe_2O_3	Al_2O_3	TiO_2	LOI
1					X	X	X	99.80	0.01	0.05	0.02	0.05
2				X	X	X	X	99.70	0.02	0.08	0.03	0.10
3				X	X	X		99.50	0.03	0.14	0.04	0.25
4		X	X					99.0	0.1	0.2	0.1	0.5
5	X	X						97.0	0.5	0.5–1.0	0.3	1.0

Applications are in the structural ceramic industry (A), foundries (B), low-grade glass (C), fine ceramics (D), high-grade glass (E), water glass, zeolites (F) and carborundum (SiC) (G). The quality per category is given in terms of minimum quartz (SiO_2) and maximum iron (Fe_2O_3), aluminium (Al_2O_3) and titanium (TiO_2) contents, and the maximum loss on ignition (LOI; %m/m). Modified from Feenstra & Mulder (2003).

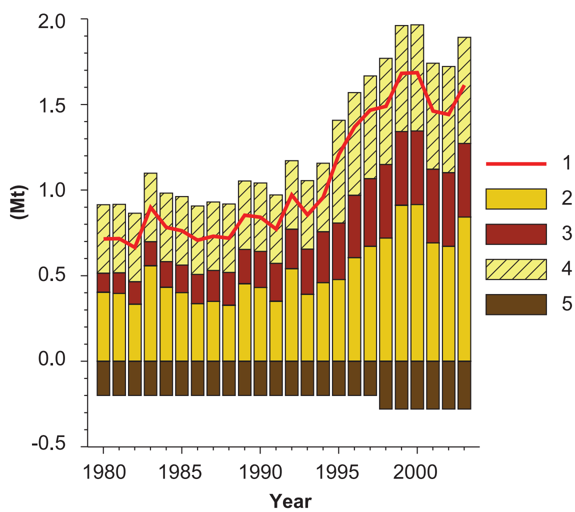

Fig. 8. The demand for silica sand in the Netherlands from 1980 to 2003 (1), and its provision by home production (2), secondary materials (3) and imports (4). The secondary substitution of silica sand is largely achieved by glass-packaging recycling (currently ~80% of the glass packaging production; Packaging Committee, 2002). High-grade silica sands are exported to Germany, Belgium and France. Export figures (5) are estimates. Sources: see Fig. 1.

Table 2. Market values of aggregates, clay (Otten et al., 2002) and shells.

Commodity	Market value (2001)
Concrete and mortar sand	8.50 €/t
Filling sand	7.00 €/m³
Silica sand	22.50 €/t
Gravel	12.50 €/t
Clay (structural ceramic industry)	10.50 €/m³
Clay (other purposes)	6.50 €/m³
Shells	26.50 €/m³

Values given include the costs of transportation to end-users within standard-sized distribution areas. The value of silica sand and, to a lesser extent, concrete and mortar sand depends on the quality: values given are averages.

RESOURCES

Currently exploited silica sands of the Miocene of southern Limburg (Breda Fm, Heksenberg Mbr), near the town of Heerlen (Fig. 9), rank among the purest in Europe. The high quartz contents are the result of a combination of a quartz-rich source material, sorting in a coastal depositional environment, and pedogenic processes. The silica sand is often intercalated with brown-coal layers, which may have acted as adsorption filters reducing the mineral content of percolating groundwater.

At current extraction rates, the permitted reserves in Limburg will last for around a further 25 years. The Dutch government has commissioned a re-assessment of silica-sand resources (Gruijters & Menkovic, 2002). The assessment included sands with qualities that do not meet the current standards of the Dutch producers, but would qualify for upgrading in other countries (e.g. Great Britain). Lesser-quality silica sands occur locally in the Kieseloolite Formation in Limburg and Noord-Brabant, the Oosterhout Formation (Lievelde Mbr) and the Peize Formation in Gelderland, and the Stramproy Formation in Noord-Brabant. The extraction and upgrading of these sands may become profitable when the currently exploited reserves have become depleted.

Gravel

APPLICATIONS AND DEMAND

Gravel is mainly used for the production of concrete. As a result of a restrictive permitting policy, national production has fallen from over 10 Mt/a in the early 1980s to 3.8 Mt/a in 2003 (Fig. 10). At a current total demand for coarse aggregates of about 32 Mt/a, the Netherlands has become largely dependent on imports of gravel and crushed-rock aggregates. The latter aggregates are used in railway beds and have substituted gravel for the production of asphalt and pre-cast concrete for a significant part. In 2000, about 13 Mt/a of gravel were imported from Germany (66%), Belgium (17%) and the British sector of the North Sea (18%). About 10 Mt/a of crushed-rock aggregates were imported from Belgium (60%), Germany

(21%), Scotland (11%), France (5%) and Norway (3%; Van der Meulen et al., 2003).

Resources

The main gravel resources are located in Limburg (Beegden Fm, Fig. 5), and have been deposited by the Meuse river from the early Pleistocene onwards. As a result of a syn-sedimentary uplift, they have been preserved in terraces, which cover the flanks of the river valley. In Gelderland, Utrecht, Noord-Brabant and Overijssel, gravel occurs mixed or intercalated within Pleistocene coarse sands, and constitutes a by-product of concrete and mortar sand production (~1 Mt/a).

There are some isolated, relatively minor surficial occurrences of gravel and gravely sand in the Dutch sector of the North Sea (Southern Bight Fm, Indefatigable Grounds Mbr). The largest of these is located in the area of the Cleaver Bank (Fig. 5, northernmost coarse-sand and gravel occurrences), which was formed as an end moraine during the last glaciation. Its estimated gravel reserve is about 30 Mt (Laban, 2002).

Clay

Applications and demand

The structural ceramic industry uses 1.5 to 2 × 10^6 m^3/a of clay, mainly for the production of bricks and roof tiles (Fig. 11). Traditionally, bricks were the most important

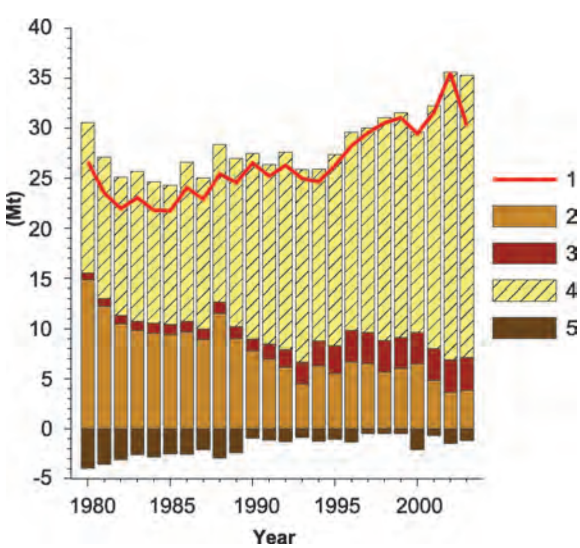

Fig. 10. The demand for coarse aggregates in the Netherlands from 1980 to 2003 (1), and its provision by home production of gravel (2), secondary materials (3) and imports of gravel and crushed rock (4). At decreasing production levels, the growing demand was met with increasing amounts of imported and secondary materials. Secondary substitution of coarse aggregates has largely been achieved by asphalt recycling. Exports (5) have decreased proportionally to the production. Sources: see Fig. 1.

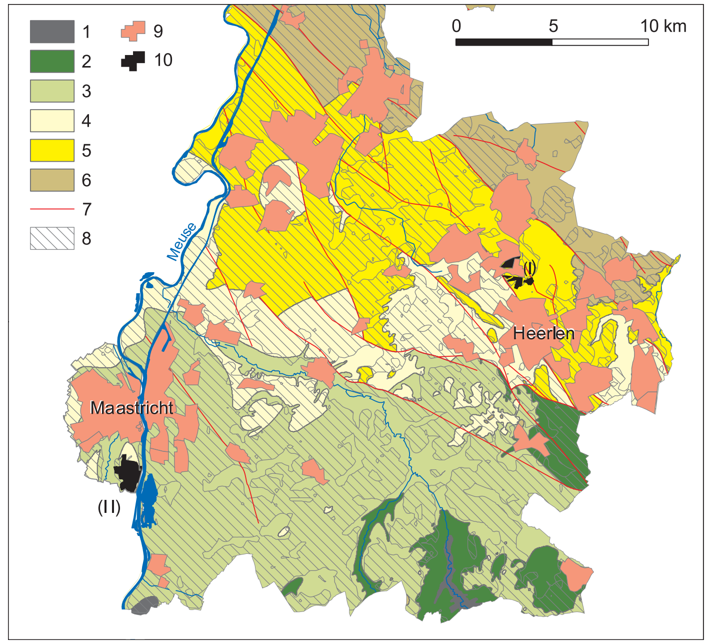

Fig. 9. Subcrop map of southern Limburg (for location see Fig. 2) showing the Carboniferous (1), Upper Cretaceous siliciclastic deposits (Aken and Vaals formations (2)), Upper Cretaceous–Lower Tertiary carbonates (chalk formations of Gulpen, Maastricht and Houthem (3)), Tertiary siliciclastic deposits (Eocene–Oligocene Tongeren and Rupel formations (4), Miocene Breda Fm (5) and Miocene–Pliocene Kieseloolite Fm (6)), and main faults (7). Hatching (8) indicates a Quaternary cover of 5 m or over; pink (9) indicates built-up areas. High-grade silica sand of the Breda Fm (Heksenberg Mbr, not shown) is extracted in three quarries (10) north of Heerlen (I). The Gulpen and Maastricht formations are the main Dutch carbonate-rock resources. The chalk quarry of ENCI (II; see text) is located south of Maastricht.

construction material in the Netherlands. Nowadays, they are mainly used for outer walls. Customer demands for certain colours or textures are important to the industry. These and other properties, such as behaviour during drying and firing, and the strength of the end product, are determined by the clay chemistry and mineralogy, and the sand and organic-matter contents (Van der Zwan, 1990; Van Wijck, 1997; Table 3). In order to obtain desired clay mixtures, producers add sands or clays from other localities to their local stocks. About 0.1 to 0.2 × 10^6 m^3/a of clay is imported for this purpose, mainly from Germany.

Clay is also used for the building and maintenance of dikes, and the covering of landfills. The material specifications for such applications are related to the desired cohesiveness and impermeability. Dike maintenance required only relatively small amounts of clay, in the order of 0.5 to 1 × 10^6 m^3/a up till the mid-1990s. At that time, the demand rose spectacularly as a consequence of a major river-dike upgrading programme, which was undertaken after two winters of exceptionally high waters and imminent flooding (Fig. 11).

Table 3. Approximate requirements for clay in the structural ceramic industry and dike maintenance.

Application	Share of grain-size fraction (% m/m)					C-org	$CaCO_3$
	Clay	< 10 μm	63–250 μm	> 250 μm	Sand	(%m/m)	(%m/m)
Roof tiles		50–57	< 20	< 40		< 3	< 25
Bricks		40–42	< 20	< 40		< 3	< 25
Extruded products		40–55	< 20	< 40		< 3	< 25
Dike maintenance	18–40				< 40	< 3	

Note that clay in the industrial definition may include fine-grained loam and silt, if sufficiently cohesive. Modified from Van der Zwan (1990).

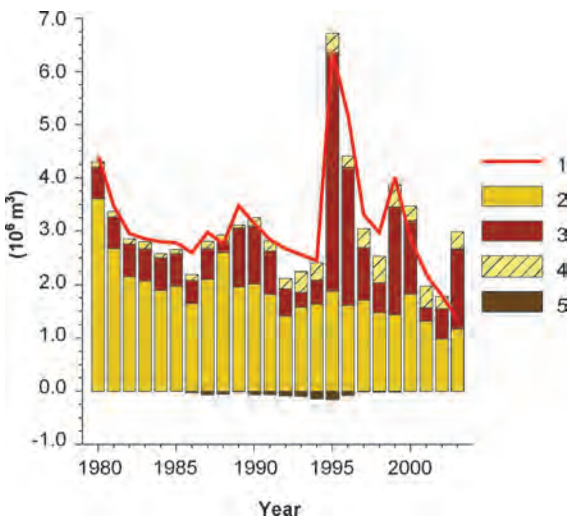

Fig. 11. The total demand for clay (1), and the production of clay for the structural ceramic industry (2) and dike building/maintenance (3) in the Netherlands from 1980 to 2003. Limited amounts of ceramic clay are imported (4) and exported (5). The peak in the production of clay for dike building in the mid-1990s is related to a river-dike upgrading programme, initiated after two phases of imminent flooding. Amounts of clay are usually given in volume rather than in weight units. Sources: see Fig. 1.

Resources

Holocene peri-marine clays in the coastal provinces (Naaldwijk Fm), and Quaternary fluviatile clays, deposited on the floodplains of the Meuse and Rhine (Beegden and Echteld formations, respectively), form the bulk of the Dutch clay resources (Fig. 12). Currently, the structural ceramic industry mainly uses the fluviatile clays; the peri-marine clays are only used in the province of Groningen. Other exploited resources include Oligocene marine clays in Gelderland (Rupel Fm), Quaternary brook deposits in Noord-Brabant, Gelderland and Overijssel (Boxtel Fm, Singraven Mbr), and Pleistocene eolian silt deposits (loess, Boxtel Fm, Schimmert Mbr) in Limburg.

Most of the ceramic clay extraction sites and permitted reserves are located in the embanked floodplains of the Rhine and Meuse (Sigmond et al., 2001). Dutch rivers were progressively embanked from approximately 1000 AD onwards. Within the confinement of the dikes, the main river channel was initially free to migrate and able to deposit sand throughout the embanked floodplains (Middelkoop, 1997). From approximately 1850 AD onwards, the river channels were fixed by a combination of groyne construction, riverbed armouring and/or strengthening of natural levees. From that time, clay became the dominant overbank deposit in the embanked floodplains, while sand transport and sedimentation became limited to the river channel. This particular clay has become the main resource for the structural ceramic industry and amounts to about two-thirds in the total consumption. For dike maintenance and the covering of landfills, locally extracted clay is used wherever possible.

Carbonates

Applications and demand

The Netherlands' only Portland-cement production plant is located near the town of Maastricht in southern Limburg (ENCI, Fig. 9). The factory operates a quarry, and currently extracts about 1.3 Mt/a of chalk (Fig. 13). The factory uses various secondary materials, such as steel slag and coal fly ash. About 50% of the cements used in the Netherlands are slag cements, containing 70 mass% of slag.

Heidelberger Cement, parent company of ENCI, has recently decided to restructure its cement-production activities in Belgium and the Netherlands; the restructuring plan includes closing down the ENCI plant. The terms and timetable of the closure are currently under negotiation: ENCI's kiln and quarry will probably be closed down on short notice, while its cement-grinding mill will be kept in operation provisionally.

About 0.3 Mt/a of chalk, limestone and dolomite is used for the production of limestone powder in Limburg and eastern Gelderland (Fig. 13). Limestone powder is mainly used in agriculture (fertilizing, fodder), as a filler in ceramic products and in asphalt, and for the desulphurisation of flue gases.

Chalk in Limburg has historically been used as dimension stone. Nowadays, only a minor amount of chalk

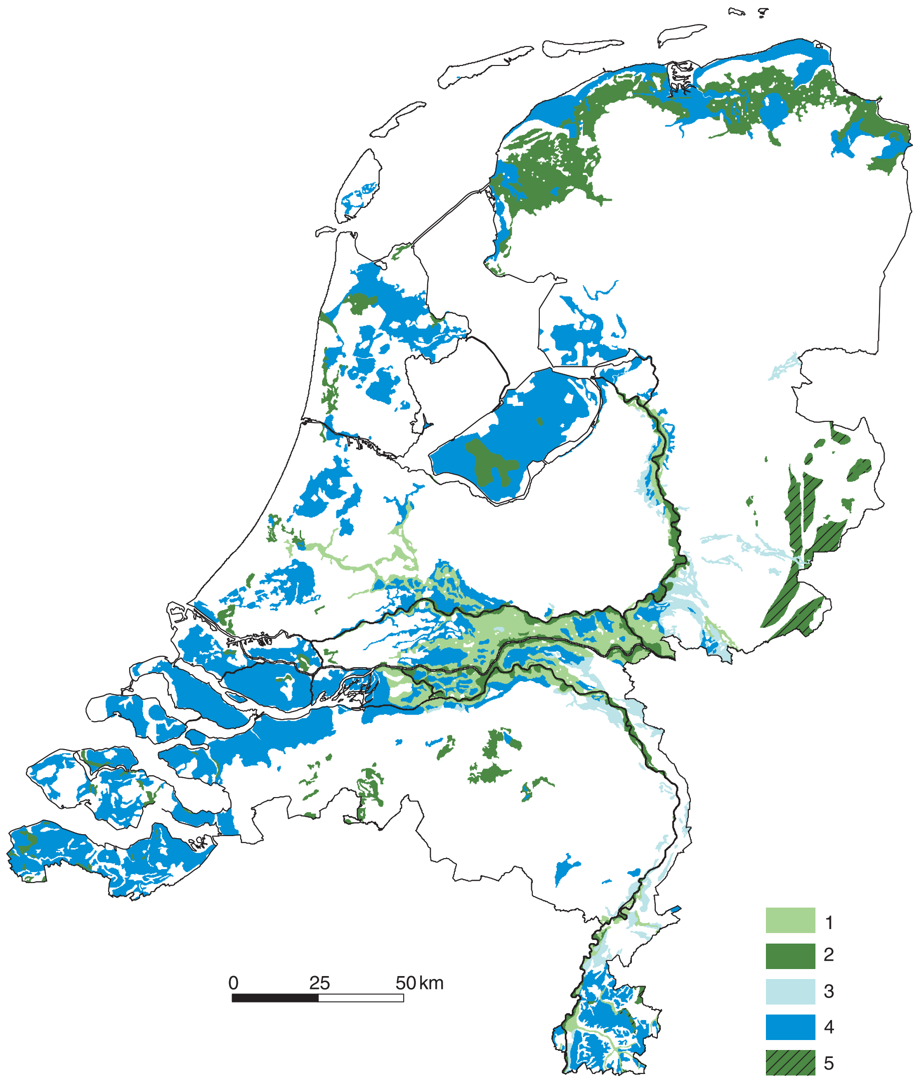

Fig. 12. The surficial occurrence of fine-grained deposits in the Netherlands: (1) Quaternary clay, loam and silt (including loess deposits of southern Limburg), thickness > 1.2 m, clay fraction < 17.5%; (2) Quaternary clay, thickness > 1.2 m, clay fraction > 17.5%; (3) Quaternary clay, thickness < 1.2 m, clay fraction < 17.5%; (4) Quaternary clay and loam, thickness > 1.2 m, clay fraction > 17.5%; (5) Tertiary clay. As dredged clays are neither applicable in construction nor in the structural ceramic industry, underwater occurrences have not been included. See also: http://www.delfstoffenonline.nl.

is quarried for this purpose, mainly for reconstruction works.

Resources

Upper Cretaceous chalk deposits in Limburg (Maastricht and Gulpen formations) form the most common pre-Tertiary, naturally exposed rocks in the Netherlands, and are the main carbonate resource (Fig. 9). The chalk has been deposited as bioclastic carbonate muds, silts and sands in a shallow-marine basin, which covered large portions of northwestern Europe during the Late Cretaceous and earliest Tertiary (Zijlstra, 1994). The chalk has a very high $CaCO_3$ content (up to 99%); the only significant siliceous component is flint, occurring in nodules or layers. The flint was used to make human tools as early as 6000 BP, and now is a by-product of limestone quarries, mainly used for road foundations, path pavements and hydraulic engineering.

Middle Triassic Muschelkalk carbonates quarried in Gelderland are an alternation of nodular limestone and dolomite, deposited in peri-marine environments.

Shells

Valves of mollusc shells are used in drains (~22%), path pavements (~20%), animal fodder (~16%), thermal and moist isolation (~13%), and in several minor applications such as foundation material and the production of lime mortar (De Graaf et al., 2000). All applications of shells,

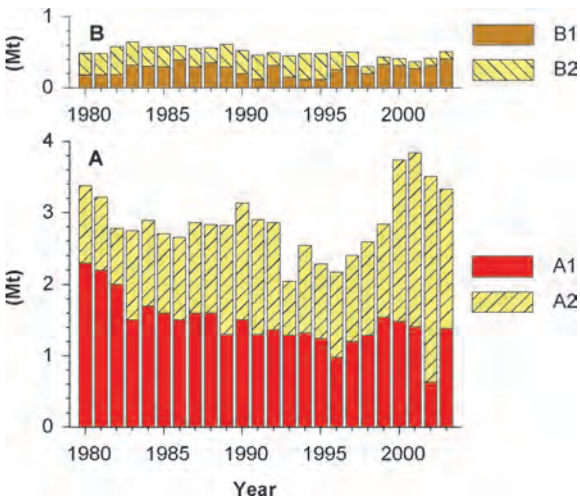

Fig. 13. Panel A: the production (A1) and imports (A2) of limestone for Portland cement in the Netherlands from 1980 to 2003. Panel B: the production (B1) and imports (B2) of limestone for other purposes (mainly the limestone milling industry). Sources: see Fig. 1.

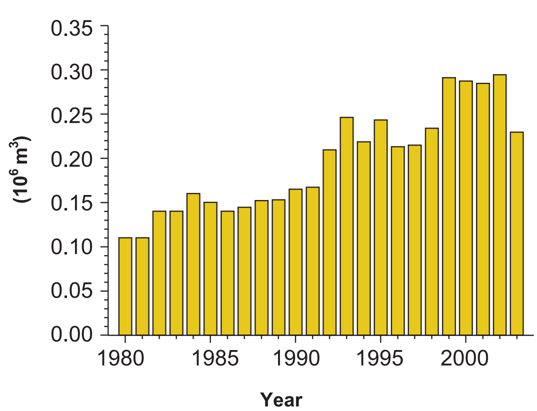

Fig. 14. The extraction of shell valves in the Netherlands from 1980 to 2003. Amounts of shell valves are usually given in volume rather than in weight units. Sources: see Fig. 1.

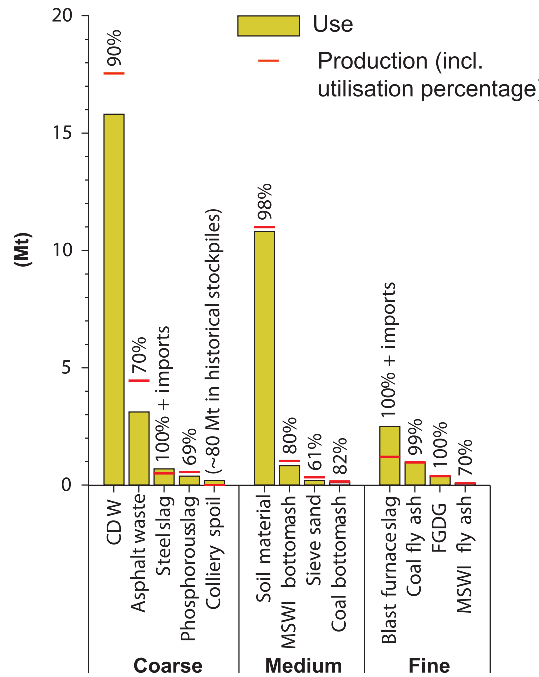

Fig. 15. The production and use of earthy and stony secondary materials in the Netherlands in 2000, in three grain-size categories: coarse, medium and fine, roughly corresponding to gravel, sand, and silt + clay, respectively. CDW = granulated construction and demolition waste, MSWI = municipal solid waste incineration, FGDG = flue-gas desulphurisation gypsum. Bottom ashes are coarse-grained incineration or combustion residues; fly ashes are fine-grained and precipitated from off-gases. Sieve sand is the sand fraction sieved from granulated construction and demolition waste. Non-utilised shares are either disposed of, or stored for later use. Sources: Van Ruiten (2002) and data at the disposal of the authors.

except lime-mortar production, are currently considered to require cockle valves (*Cerastoderma edule*), which endure intensive use and mechanical processing. When crushed, the combination of concentric growth increments and radial ribs results in approximately equidimensional fragments: a useful property for fodder production and path pavements.

The current demand for shells is about 300 000 m^3/a. In the 1990s, the extraction has risen from about 180 000 to 290 000 m^3/a (Fig. 14). About 70% of this amount is extracted from Holocene deposits of the Waddenzee and its inlets (Naaldwijk Fm). The remaining 30% is mainly extracted from the Pleistocene of the Westerschelde estuary and its prodelta. For reasons of sustainability, the annually permitted volumes in these areas are limited to the estimated average yearly growth increments of the exploitable shell stocks (cf. Beukema & Cadée, 1999). About 10 000 m^3/a of cockle shells are obtained from seafood producers.

Secondary materials

The utilisation of various waste materials and by-products has resulted in a reduction of the use of the above natural materials in most of their applications. The share of secondary materials in the total provision has risen from ~7 Mt/a (6%) in 1980 to ~32 Mt/a (20%) in the early 2000s (Fig. 1). Coarse and medium-grained secondary aggregates are mainly used in construction (Fig. 15). Granulated construction and demolition waste, steel slag and phosphorous slag have become widely used road-foundation materials. Asphalt waste is recycled into new asphalt. Colliery spoil is used for hydraulic engineering, and as fill or pavement material. Medium-grained materials mostly serve as secondary alternatives for filling sand.

Fine-grained industrial, waste-processing and energy-production by-products have become well-accepted raw materials in the building-materials industry. Ground granulated blast-furnace slag is used in the cement industry. Municipal solid-waste incineration fly ash is used as filler; coal fly ash is used for both purposes. Flue-gas desulphurisation gypsum has become a widely used secondary alternative for primary gypsum.

At present, land-filling with, or disposing of waste materials is taxed, and banned for several recyclable materials. Nonetheless, the utilisation of secondary materials has not yet reached its full potential. Currently, policy and research are directed towards using dredging sludge from waterway and harbour maintenance (Ringeling & Rienks, 2002). Depending on dredging investments, the annual supply in the coming years will vary between about 7 and 15 × 10^6 m^3 (Broers et al., 2002). Up to ~4 × 10^6 m^3/a of this amount could be used in construction as an alternative for natural clay or filling sand.

Some historically exploited resources

Peat (Nieuwkoop Fm and Boxtel Fm, Singraven Mbr), deposited in Quaternary paralic, fluviatile and lacustrine environments, has been extracted from medieval times till the mid 20th century. Turf (dried peat) has historically been used as fuel (Van Bergen et al., this volume). In the southwest of the medieval Netherlands, paralic peat has also been extracted and burned to produce salt from the ashes. The current demand for turf, for the production of horticultural compost etc., is almost completely met by import, mainly from Germany. Net imports in 2002 were 0.6 Mt (Statistics Netherlands (CBS), Voorburg).

In the central and southeastern parts of the country, iron has been extracted from bog ore, iron (hydr)oxide concretions (mainly from Pleistocene ice-pushed ridges in Gelderland), and from siderite nodules (e.g. from the Upper Carboniferous of southernmost Limburg; Laban et al., 1988; Joosten, 2004). Iron production started in prehistoric times and reached a high in medieval times. Nowadays, imported iron ore is used to produce steel. Net imports were 7.4 Mt in 2002 (Statistics Netherlands (CBS), Voorburg).

Solid sedimentary rocks such as the aforementioned carbonates and Carboniferous sandstones occurring in Limburg, have historically been used as dimension stone. Nowadays the Netherlands depends almost completely on imports of dimension stone, mainly from Belgium and Germany. Net imports of uncrushed carbonates, sandstones and crystalline rocks amounted to 3.2 Mt in 2002 (Statistics Netherlands (CBS), Voorburg).

Pleistocene subglacial tills ('keileem', Drente Fm, Gieten Mbr), occurring in the northern and eastern parts of the country, have been used for dike building, in particular for the 'Afsluitdijk', which separates the IJsselmeer lake from the Waddenzee (Fig. 2). Its construction, in 1927–1932, required 13.5 × 10^6 m^3 of till and 23 × 10^6 m^3 of sand.

In eastern Gelderland and Overijssel, Oligocene phosphorite nodules occur in the shallow subsurface, either as a primary constituent of marine deposits (Rupel Fm), or reworked in Pleistocene moraines (Laban, 1988). These nodules were processed into fertiliser during the First and Second World Wars, when all imports were restricted. The current demand for phosphate in chemical fertilisers is about 50 kt/a (Agricultural-Economics Research Institute (LEI), Wageningen).

Diatomites are mainly used in various industrial filters, as grinding and polishing material, as a reactive silica source, and for the production of paint, matches and dynamite (Vos & De Wolf, 1988). The only known Dutch exploitable resource, a Holocene lake deposit near Renkum, southern Gelderland, was virtually depleted in a 40-year period around the turn of the 19th century (Laban, 1987). Current import figures for diatomite are not available, but are expected to be low.

Exploration and production

Exploration for aggregates

Exploration by Dutch sand and gravel extraction companies is aimed at estimating the tonnage of specific mixtures of sand and/or gravel that can be produced at a potential site. This requires a high-density drilling campaign (of up to one drilling per 10 000 m^2) and grain-size analyses on a sample per metre drilled. Within such a dense drilling grid, additional geophysical techniques are considered to have limited added value, and are seldom used for commercial exploration. Another important exploration objective is to identify cohesive intercalations of clay or peat which may inhibit exploitation with dredging devices. The exploration depths vary between approximately 20 and 40 m, depending on the geological situation and the local policy concerning the depth of extraction pits.

The exploration results are interpreted in terms of the estimated costs of the removal of the overburden and the subsequent extraction and processing of the commodity. These costs, together with those for obtaining the concession area, the required permits, transportation etc., determine the price of the commodity and the viability of the project.

The Dutch government occasionally commissions aggregates exploration studies for inventory or planning purposes, or in connection with large-scale hydraulic-engineering projects. Recent investigations include the exploration for concrete and mortar sand in the North Sea, and the combination of hydraulic measures and sand or gravel extraction along the major rivers. In the initial stages, such research heavily depends on existing drilling data: the Netherlands Institute of Applied Geoscience TNO, the national geological survey, maintains a database (DINO) of some 380 000 digital borehole descriptions for such purposes. In the data-gathering stage, the scale of the projects precludes the high drilling density of commercial exploration and calls for geophysical, mainly seismic techniques for mapping relevant sediment units.

While exploration for sand and gravel requires more or less standard geological and geophysical techniques, exploration for ceramic clay and silica sand may involve various types of specialised chemical and mineralogical analyses (Brijsse & Casier, 1997; Walda, 2000). Their interpretation requires the context of production processes, and is not further discussed here.

Extraction of aggregates and clay

Most of the sand and gravel produced in the Netherlands is dredged from extraction pits or in open water. Dry extraction, with equipment like draglines, bulldozers and shovels, is used for clay and only occasionally for sand and gravel (e.g. silica sand in southern Limburg).

Both mechanical (digging), and hydraulic (suction) processes are used for wet extraction. Gravel is usually dredged mechanically, suction being less suited for reasons of wear. Suction dredgers are preferred for sand extraction because of their high production rates and the possibility to use pipelines for transport.

The bucket dredger is one of the oldest types of dredgers. It features a chain of buckets, supported by a ladder, which fill while scraping the bottom (Fig. 16a). Bucket dredgers are still in use for coarse-sand and gravel extraction in Limburg. Their main advantage is the possibility to dredge very accurately. Disadvantages are the limitations in depth (down to 20 m) and the noisy operation. Since 1980 a few bucket dredgers have been modernised to reduce the noise and accommodate depths down to 35 m.

The wire grab dredger is a versatile excavator suitable for depths down to 100 m (Fig. 16b). For sand and gravel extraction, it has largely been replaced by the backhoe dredger which has a higher production rate and allows for more accurate dredging. Backhoe dredgers are based on a digging action towards the machine and can reach depths down to 15 m (Fig. 16c).

A stationary suction dredger is the most common device to extract sand from pits (Fig. 16d). Suction causes breaching, which generates a flow of sand towards the suction pipe. The process may be disturbed by cohesive sediments, like peat or clay. Such materials can be loosened or

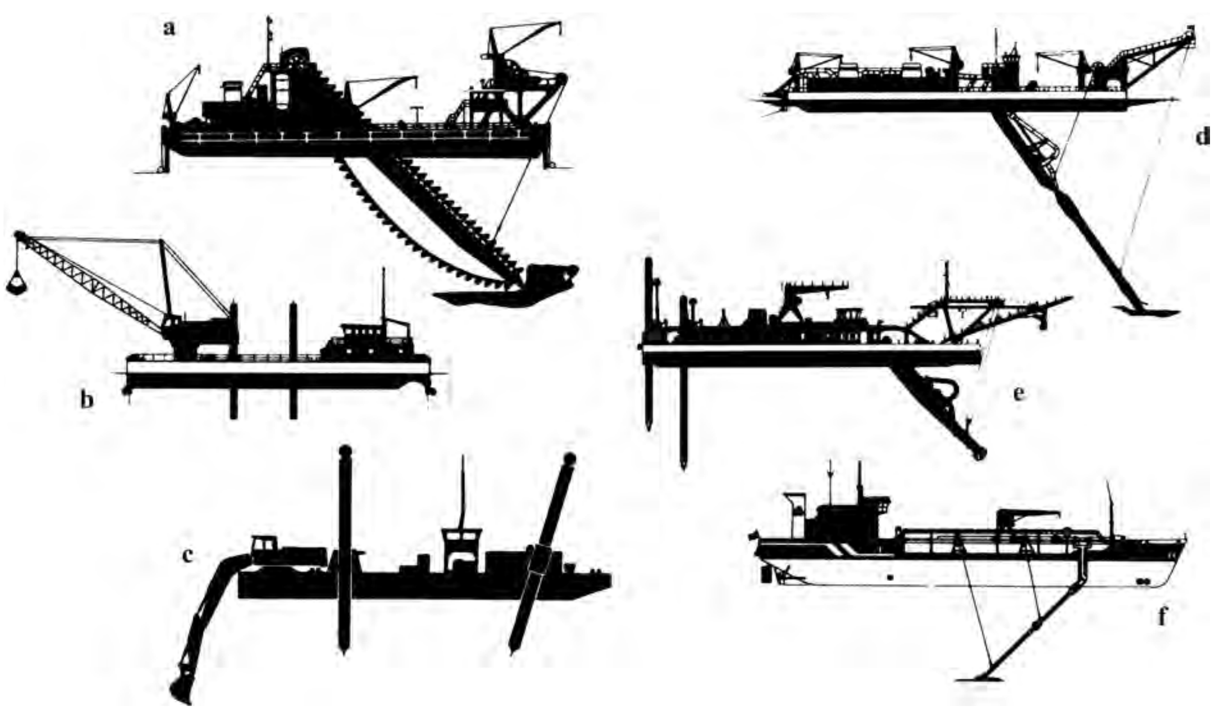

Fig. 16. Dredging vessels used for sand and gravel extraction, with an indication of dredging rates: a) bucket dredger (200–1000 m^3/h); b) wire grab dredger (100–300 m^3/h); c) backhoe dredger (200–500 m^3/h); d) stationary suction dredger (500–4000 m^3/h); e) cutter suction dredger (1000–3000 m^3/h); f) trailing suction hopper dredger (500–7000 m^3/h). Modified from Donze (1990).

Fig. 17. Floating classification unit for sand and gravel. Sand and gravel enters at the left-hand side for treatment in clarification tanks and bucket-wheel separators. In the mid section, various size classes of gravel are separated by sieves. The sand fraction is classified by sieves, hydrocyclones and fluid-bed separators at the right-hand side. Photo courtesy of Smals BV, Herten.

penetrated by a suction pipe equipped with water jets. The standard suction dredger is limited to a depth of 30 m, but equipment capable of dredging down to 70 m is available.

A cutter suction dredger is capable of dredging cohesive materials with a combination of a cutter head and hydraulic transport (Fig. 16e). Cutter suction dredgers are used for the construction and maintenance of harbours and navigation channels, and occasionally for the extraction of sand and gravel.

The trailing suction hopper dredger is a self-propelled vessel which scrapes a thin layer from the bottom, and stores the dredged material in its hold (Fig. 16f). This device can only be used in deep, open waters. Advantages are the high dredging accuracy, high production rates and its capability to operate in rough waters. Hoppers have been built with carrying capacities ranging from about 2000 to 25 000 m^3, and operating depths from 20 down to 130 m. Filling sand from the North Sea is extracted with medium-sized hoppers.

After extraction, gravel and sand (except filling sand) are classified by sieving and hydrocyclonage. Floating classification units can be mounted on the dredging vessel, or on a separate pontoon (Fig. 17). Larger or more sophisticated plants are either mounted next to sand or gravel pits, or at central locations where sand from several extraction sites is processed (Fig. 18).

Societal aspects

Economy

About one thousand Dutch companies are involved in the extraction and trade of quarry products. Six hundred of these are held by about forty concerns. There are about eighty independent small and medium-sized extraction companies and three hundred trade companies. Some of the larger concerns are vertically integrated and own, for example, mineral-extraction operations supplying their building-materials factories or construction projects. The smaller independent companies are often family-owned. The sector as a whole is small in terms of the value of production and contribution to the gross domestic product (Table 4). Its financial results, 35% of the value of production, are well above the Dutch average of ~20% (Otten et al., 2002). Multinationals play a significant role in the building-materials industry, but have limited interests in the extraction of minerals, except for silica sand and chalk. The Dutch dredging companies involved in the extraction of filling sand from the North Sea, are global players.

In recent years, internationalisation has been one of the most relevant trends in the sector (Van der Meulen et al., 2003; Wagner et al., 2005). At more or less fixed production possibilities in the country and an ever-increasing reliance on imports (Fig. 19), Dutch mineral-extraction com-

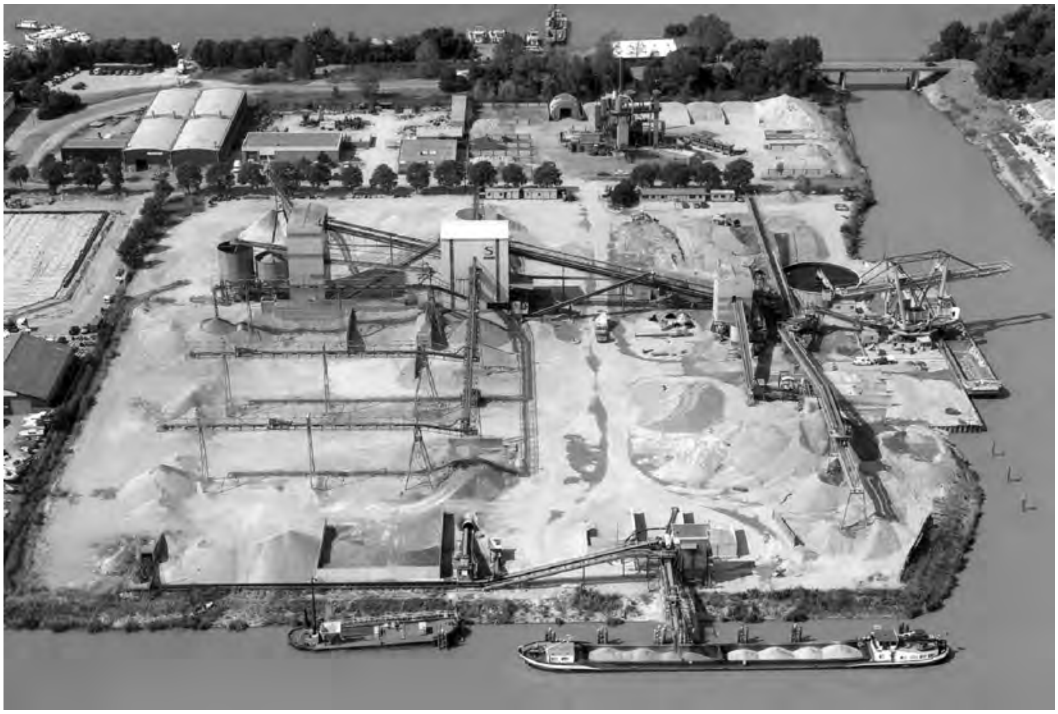

Fig. 18. Large classification plant for coarse sand and gravel in the southern Netherlands. Sand is delivered by ship at the right-hand side. After classification it is stored in piles per size class on the left-hand side. Photo courtesy of Smals BV, Herten.

Table 4. Economic figures (1999) for the Dutch quarry-products sector, and two related sectors.

Sector	Value (M€)			Contribution to GDP (373 664 M€)	Financial result	
	Production	Consumption	Added		(M€)	(% of value of production)
Quarry products	871	364	438	0.1%	309	35
Glass, ceramic, cement and limestone products	5 798	3 076	2 111	0.6%	1074	19
Building and construction	54 052	36 627	17 426	4.7%	4003	7

GDP = gross domestic product. Data from Statistics Netherlands (CBS, Voorburg) and Otten et al. (2002).

panies have expanded abroad to maintain or enlarge market shares and turnover. A significant number of companies on the Belgian side of the Meuse river and in the German border area east of Gelderland and Limburg are Dutch-owned.

Legislation

The exploitation of surface mineral resources is governed by the revised Mineral Extraction Law ('Ontgrondingenwet'; Anonymous, 1996a). Extraction permits are issued by the provincial administrations, and by the Ministry of Transport, Public Works and Water Management where State Waters ('Rijkswateren') are concerned. Areas where quarrying is considered acceptable are outlined in mineral-extraction plans, which are issued as a part of, or in accordance with regional development plans ('streekplannen' or 'omgevingsplannen'). Tree-cutting, noise, turbulence and other environmental impacts associated with extraction may require additional permits from local, e.g. municipal authorities. Application procedures tend to be time-consuming, and may take well over five years.

Policy

Quarrying has been challenged by an ever growing societal resistance since the 1960s. As a result, provincial administrations became increasingly reluctant to grant extraction permits, and by the mid-1970s supplies of gravel, concrete and mortar sand, and clay for the structural ceramic industry became a matter of national concern (Ike, 2000; Van der Meulen, 2005a). The problem was taken up by the Ministry of Transport, Public Works and Water Management, which started preparing for mineral planning guidance on a national level. As a first step, the Ministry and the provincial administrations agreed on a set of assignments ('taakstellingen', i.e. amounts of product for which extraction permits are to be granted) for the extraction of gravel, concrete and mortar sand, and clay during the 1980s.

The next step, in the late 1980s, was to draft the first national policy plan on surface mineral extraction. It stated that the exploitation of surface mineral resources requires planning and co-ordination, and better embedding in other policy fields, especially spatial planning (Anonymous, 1989). Policy from then on aimed at sustainable exploitation of the surface mineral resources to meet the demand for construction and building materials, at an economical use of materials and maximum use of renewable and secondary materials (Anonymous, 1996b, 2001).

Production assignments, though abandoned for clay in 1989, remained the primary policy instrument in mineral planning. Up to 1998, the production levels of concrete and mortar sand were kept in approximate balance with the national demand. For 1999–2008, the authorities agreed on an underproduction of about 20% in order to stimulate the use of secondary alternatives (Anonymous, 2001). Instead, it seems to have resulted in increased imports (Fig. 4). In 1989, the production levels of gravel in Limburg, which include a share for the supply of national markets, were allowed to become eventually reduced to a level of provincial self-supply. The final tens of millions of tonnes of gravel for national supplies will be extracted in conjunction with widening and deepening of the river Meuse between 2005 and 2017.

Generally, surface mineral extraction is considered more acceptable if it is combined with other activities (e.g. Van der Meulen et al., 2004). The coupling of aggregates extraction and river widening is also envisaged for the Rhine. As a result of an agreement between the World Wide Fund for Nature (WWF) and the Dutch organisation of brick producers, the extraction of brick clay has become coupled with nature development.

The national government recently decided to take on a lesser role in mineral supplies. Mineral planning became embedded in the national policy on spatial planning ('Nota Ruimte', Anonymous, 2004) and is no longer outlined in a specific policy plan. The policy goal is to stimulate surface mineral extraction in a socially responsible manner. Mineral-extraction sites should, where possible, contribute to the realisation of societally desired functions such as nature, recreation, lakeside housing and water

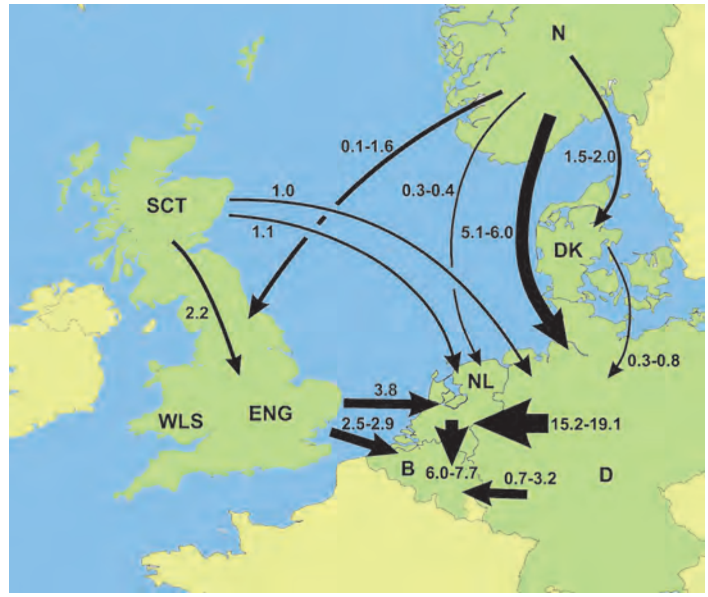

Fig. 19. The aggregates trade between the Netherlands (NL), Germany (D), Belgium (B), Scotland (SCT), England / Wales (ENG / WLS), Denmark (DK) and Norway (N) in 2000 (reprinted from Van der Meulen et al., 2003, with permission of the publisher). Arrows indicate directions of net supply, i.e. the positive difference between exports and imports, given in Mt. Net supplies below 0.5 Mt are not shown. The Netherlands and Belgium clearly are net importing countries.

management. To this end, a so-called building raw materials assessment ('bouwgrondstoffentoets') has been introduced (for a discussion see Van der Meulen, 2005b). Initiators of spatial plans have to evaluate (i) the effects on mineral supplies, (ii) the accessibility of resources of scarce minerals for future generations, and (iii) the possibility of embedding mineral extraction into projects which have other primary goals. The assessment does not apply to plans for already built-up areas. Policy elements regulating the market, especially the production assignments, have been abandoned, but standing agreements for 1999–2008 will be fulfilled. The policy changes imply that the Mineral Extraction Law will have to be reconsidered. An option is to abolish it and to transfer indispensable rules to other laws or decrees.

Future developments

Figure 20 shows scenarios for the demand for aggregates, clay and carbonates for 2003-2015, based on models for the overall long-term macro-economic development of the Netherlands. After recovery from the economic dip of the early 2000s, a gradual average increase of the demand is predicted, which is expected to be partly met with secondary aggregates.

Deregulation of the aggregates production will especially affect supplies of concrete and mortar sand. Home

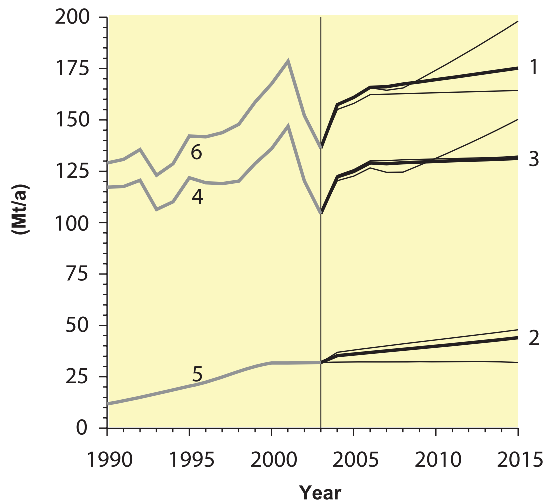

Fig. 20. The forecasted Dutch demand for aggregates, clay, and carbonates for 2003–2015 (1), utilisation of secondary materials (2), and demand for primary materials (3). Thick lines correspond to a medium scenario for the overall long-term macro-economic development. Thin lines correspond to low and high scenarios. Forecast models have been prepared by JWB for Anonymous (2001). For reference, the use of primary (4) and secondary materials (5), and the total consumption (6) are shown for 1990–2003.

production levels may well fall back, and it is as yet unclear how this will be compensated. Theoretically, up to ~25% of all aggregates can be substituted by secondary materials, especially by granulated construction and demolition waste which is now mainly used in road foundations (Hendriks & Pietersen, 2000). However, the building sector is reluctant to use these alternatives for traditional materials. Consequently, it is most likely that the Netherlands will become increasingly dependent on imports of aggregates.

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Underground storage and sequestration

C.F.M. Bos

ABSTRACT

In the Netherlands, underground storage of natural gas (UGS) is the only form of underground energy storage currently operational. It serves to meet peak demand for gas when producing fields are unable to deliver at the required capacity. Until recently, the reservoir pressure of the giant Groningen field was sufficient to balance supply and demand instantaneously, even under extreme winter conditions. At present this field can no longer guarantee supply–demand equilibrium under all circumstances. Therefore, three smaller gas fields have been converted to UGS operation. Together they can furnish a maximum gas rate of ca. 140×10^6 m^3/d and cycle (i.e. inject and reproduce) ca. 5×10^9 m^3/a. One additional facility in a salt cavern is being planned. Further installations may become necessary. A potential new form of underground energy storage is compressed-air energy storage (CAES) as it may help to counteract the variability in electricity supply due to the planned development of large-scale offshore wind energy. The feasibility of CAES in salt caverns is being investigated. The objective of underground sequestration of residue gases such as CO_2 is to avoid their emission into the atmosphere. One field pilot test is ongoing: a relatively small amount of CO_2 is injected into a depleted offshore gas reservoir. Other forms of sequestration have not yet reached the field testing stage. Examples are CO_2 sequestration in coal seams, potentially stimulating the recovery of methane adsorbed on the coal, and CO_2 flooding of oil or gas reservoirs, potentially resulting in enhanced oil or gas recovery. The disposal of toxic and radioactive solid waste was investigated in the recent past, but to date no field tests have been carried out.

Keywords: Netherlands, underground gas storage, underground CO_2 sequestration, compressed-air energy storage, underground waste disposal

Introduction

An important recent function of the Dutch subsurface is its current and potential use as a storage medium. One type of subsurface storage is the pore volume in rocks such as sandstones, where the pores are filled with water, natural gas or crude oil. That pore volume is enormous and part of it may function economically as a container for various types of fluids, i.e. liquids or gases, that are injected from the surface into boreholes penetrating the porous medium. These fluids may be stored for short to medium-term durations or (semi-)permanently. Another type of subsurface storage media are large voids, e.g. caverns that have been leached in rock salt. The abundance of salt layers and diapirs in the northern and eastern Netherlands allows large and stable voids to be developed by solution mining. Storage of fluids in such caverns may be more attractive than storage in porous media. The latter provide more flow resistance and, hence, are economically less efficient when high flow capacities are required.

This chapter briefly reviews the following aspects of underground storage in the Netherlands:

- the temporary storage of natural gas, either in porous sediments (sandstones) or in salt caverns, as an option to meet peak demand;
- the (semi-)permanent storage or ‘sequestration’ of CO_2 after having been captured at industrial sites or power plants and after having been transported through pipelines to the sequestration site;
- the temporary storage of compressed air in salt caverns as a potential option to balance instantaneously electricity demand and supply, thereby avoiding disruptive voltage fluctuations in the power grid due to the planned large-scale development of offshore wind farms;
- earlier plans for underground storage of energy by means of ‘pumped-hydro’;
- the past research on toxic and radioactive solid waste disposal.

An earlier review of the subject was published by Visser (1987).

Underground storage of natural gas

Objectives

The objective of storing gas underground is to meet peak demand when this can no longer be satisfied by production from regular gas fields. This situation may arise if the pressure in a partially depleted gas reservoir has decreased to a level at which the gas cannot be supplied to the market sufficiently rapidly. Until recently, the giant Groningen field (estimated ultimate recovery 2700×10^9 m^3) in the north of the country acted as a ‘swing producer’, because it could meet the so-called ‘swing’, i.e. the seasonal and daily or hourly fluctuations in demand, of the entire

Netherlands as well as of the buyers in other countries. However, since more than half of the initial reserves in the Groningen field has been produced, the reservoir pressure has fallen proportionately, and currently the daily or hourly peak demand has to be supported by other facilities. The continuing depletion of the Groningen field will create the need for additional underground gas-storage facilities to meet the national peak demand. Moreover, in view of the favourable location of the Netherlands in the western European gas infrastructure, additional facilities can be important for providing flexibility services to the international market in the medium to long term (UN Economic Commission for Europe, 1999; International Energy Agency, 1995).

Mode of operation

Typical terminology used in underground gas storage (UGS) includes the terms cushion-gas volume, working volume, injection capacity and withdrawal or 'send-out' capacity. Although the cushion-gas volume and the working volume together make up a single volume of underground gas, only the working volume is withdrawn and injected cyclically. The cushion-gas volume represents a once-only investment, and only serves to maintain a minimum reservoir pressure. As the working volume is withdrawn in winter, when demand is high, the reservoir pressure falls, together with the send-out capacity. However, in summer, when demand is low, new gas is injected to replace the produced working volume. Consequently, the reservoir pressure rises and the injection capacity gradually decreases. Key characteristics of a UGS facility are the maximum injection and send-out capacities at the start and the end of the working-volume cycle respectively.

Possible types of underground storages

For relatively large working volumes, the preferred option in the Netherlands is to store gas in a (partially) *depleted gas field*, i.e. in a porous medium such as a sandstone. The advantage is that crucial information on such fields is already available (seismic, geological and well information, production history), and that their long-term sealing capacity has been proven by nature. Compared to salt caverns, the disadvantage is that the volumes of gas fields may be too large for the design criteria, and that the porous nature of the reservoir implies significantly more internal flow resistance. If the design criteria specify a relatively small working volume as well as frequent and fast changes from injection mode to maximum send-out capacity, a depleted gas field may be unsuitable and a *salt cavern* may be preferred. Such a cavern is created by dissolving the salt surrounding a well bore by circulating fresh water (Geluk et al., this volume). The maximum volume of the cavern is limited by its roof stability and by the intrinsic plastic flow behaviour of salt: under differential stress, such as at the cavern's walls, salt tends to behave like a viscous fluid. Salt flow could cause the cavern to cave in slightly, especially at larger depths (say over 1500 m). The storage facility's cyclic pressure fluctuations may exacerbate this stability problem by weakening the mechanical properties of the salt (fatigue). Salt caverns are therefore only suitable for storing relatively small working volumes of gas at limited depths. A practical upper limit for the volume would be ca. 500 000 m^3. The capacity can be increased by constructing multiple caverns and linking them through the surface facilities. In this case, a cavern spacing of a few hundred metres should be maintained in order to avoid stability problems. Moreover, potassium and magnesium salts, that tend to 'squeeze' into a void more easily than sodium salt, should be avoided.

The option of constructing a facility in an *aquifer* is unlikely to be realised unless the other two options are impractical or not available. The sealing capacity of an aquifer's overlying layer (e.g. shale) would have to be assessed first, and this type of study is expensive and time-consuming. Moreover, there is generally much less information available on aquifers than on gas fields that have been producing for years. To prevent any leakage of gas, monitoring the gas saturation by means of 4D seismic surveys will usually be essential and increase the costs. Finally, the injectivity of wells in aquifers is often problematic.

Existing and planned facilities

At present there are three underground gas-storage facilities operational in the north of the Netherlands (Table 1). They are located in the Norg (province of Drenthe), Grijpskerk (Groningen) and Alkmaar (Noord-Holland) gas fields. To meet both domestic and international peak demand, additional facilities are expected to be constructed in the near future. At the time of writing (November 2005), the following new underground gas-storage facilities are planned:

– Essent is working on plans to convert the Waalwijk gas field (Noord-Brabant) into a gas-storage facility. The sandstone reservoir is part of the Lower Germanic Trias Group (Triassic).
– A joint venture of Nuon, Gasunie and AkzoNobel recently started the construction of new caverns in the Permian Zechstein salt near Zuidwending (Groningen) for use as gas storage. The start of operation is planned around 2008-2010.
– Just across the German border, near Epe, Essent is planning to use salt caverns for delivering swing capacity to the Dutch market.

Underground sequestration of CO_2

CO_2 sequestration involves the post-combustion capturing at power plants of the CO_2 that is normally released

Table 1. Characteristics of the three underground gas-storage facilities currently operational in the Netherlands.

UGS facility	Norg (also named Langelo)	Grijpskerk	Alkmaar
Reservoir formation	Rotliegend	Rotliegend	Zechstein
Depth of reservoir (m sub-sea; approx.)	3 000	3 200	2000
Max. daily withdrawal (10^6 m^3/d)	51	56	36
Min. daily withdrawal (10^6 m^3/d)	4	4	2.4
Max. daily injection (10^6 m^3/d)	12	12	3.6
Min. daily injection (10^6 m^3/d)	10	10	?
Total gas volume (10^6 m^3)	28 000	13 100	3600
Working gas volume (10^6 m^3)	3 000	1 500	500
Cushion gas volume (10^6 m^3)	25 000	11 600	3 100
Ratio working/total gas (%)	10.7	11.5	13.9
Duration of withdrawal (days)	69	29	14

into the atmosphere. Instead of venting the flue gas into the air, the CO_2 is separated from the flue gas and injected into the subsurface. All signatory countries of the Kyoto protocol, including the Netherlands, are committed to reducing their CO_2 emissions into the atmosphere (website climnet.org). To comply with the protocol on the medium to longer term, the capture and underground storage of CO_2 will probably be necessary until the CO_2 emission problem can be solved by energy savings, and/or by the application of nuclear power and sustainable forms of energy (notably biomass and wind). The CO_2 injected into the subsurface will have to remain stored for hundreds of years. Therefore, the long-term sealing behaviour of the cap rock of the reservoir should be well understood in order to prevent leakage into the soil or atmosphere (websites co2geonet; co2captureandstorage).

Like in underground gas (i.e. methane) storage, CO_2 can be sequestered in depleted oil or gas fields, aquifers and abandoned mines. In all cases the storage medium must be sealed both laterally (e.g. by an impervious fault) and vertically (e.g. by an overlying layer of shale or rock salt). CO_2 can also be sequestered in coal seams. This can be combined with the production of coalbed methane: the CO_2 that is injected into the coal seams will preferentially adsorb on the coal, thereby replacing and releasing the originally adsorbed methane (Van Bergen et al., this volume). Theoretically, a win-win situation may be created when CO_2 sequestration is combined with the simultaneous production of methane gas. This process, called enhanced coalbed-methane recovery (ECBM), could become a viable option in the Netherlands. Spurred by an ongoing ECBM project in Poland (Silesia), some preliminary investigations into the applicability of ECBM in the province of Limburg have been carried out (Van Bergen et al., 2004).

Other win-win situations are also possible. CO_2 can be used in the upstream petroleum industry for the enhanced recovery of oil and gas. When CO_2 is injected into oil-bearing strata at sufficiently high pressure, it becomes miscible with the oil. This reduces the capillary forces that trap part of the oil in the pores of the reservoir rock, resulting in less oil being left behind in the reservoir. Moreover, the viscosity of the oil will be reduced, resulting in higher production rates. This type of oil recovery has been standard practice for many years in some fields in Texas, USA (Holtz et al., 1999). The injection of CO_2 into a gas-bearing reservoir provides pressure support, allowing a faster production of the natural gas. However, field tests for enhanced gas recovery by CO_2 injection have not yet been implemented, because it is feared that the injected CO_2 will mix with the natural gas in the reservoir thereby contaminating the production stream and resulting in corrosion of the installations (nevertheless, a test in the Netherlands is being considered; see below).

CO_2 sequestration in salt caverns seems unpractical; safe storage may perhaps even be impossible. The maximum volume per cavern (ca. 500 000 m^3) would allow only ca. 0.35 Mt of super-critical CO_2 to be injected. This is equivalent to the storage capacity of a very small gas field of ca. 0.2×10^9 m^3. More importantly, however, due to the tendency of salt to flow, the pressure of the CO_2 bubble will eventually increase to the lithostatic pressure. Whether under such circumstances the CO_2 will remain contained or permeate out of the salt is unknown.

A first field test for underground sequestration of CO_2 has recently started offshore the Netherlands by Gaz de France. CO_2 is currently being injected into the depleted Rotliegend reservoir of the K12-B gas field. Several other pilot projects for CO_2 sequestration are ongoing in other countries, e.g. in Norway and Germany.

Other tests in the Netherlands are in the planning phase. For example, a plan for a Zero-Emission Power Plant ('ZEPP') near Drachten (Friesland) has been submitted to the authorities, with the post-combustion CO_2 being stripped from the flue gas and subsequently injected into a producing gas field in the nearby Akkrum concession.

The additional pressure support from the CO_2 injection may enhance the ultimate recovery of the gas.

Finally, the ongoing CRUST project (CO_2 Re-use through Underground STorage) is considering how to devise a scheme that supports market players in the realisation of an underground CO_2 storage that allows the recovery and reuse of CO_2 in the future. The CRUST project is being supported by the Dutch authorities.

Underground storage of compressed air

Objectives

Compressed-air energy storage (CAES) in subsurface salt caverns in the Netherlands is currently being considered as a possible solution to address expected balancing problems in the national electricity grid resulting from the planned increase in wind power. The basic idea of this type of energy storage is to transfer off-peak energy produced by nuclear, natural-gas or coal-fired power plants to the high-demand periods, using only a fraction of the gas or oil that would be used by standard peaking machines, such as a conventional gas turbine. The off-peak energy produced by wind turbines in offshore wind farms could also be transferred in this way. The higher the volatility in the kilowatt hour price, the more compelling the business case for storing electrical energy. The price volatility in the Netherlands already being one of the highest in the European Union, large-scale production of wind energy such as currently planned in the Netherlands will only exacerbate this, increasing the need for some form of massive storage. Also from the point of view of security of supply, large-scale wind energy would impose too much 'stress' on the national power grid. During times of strong winds and low demand it may become difficult to balance demand and supply, thereby potentially leading to undesired side-effects, including the local disruption of power supply. Moreover, when the wind is blowing irregularly, the start-up time required to switch on stand-by conventional power units may be too long, also potentially resulting in balancing problems.

Mode of operation

Compressed-air energy storage works as follows. During off-peak demand periods, a compressor consumes electrical power to pump air into the underground storage. Later, during peak-demand periods, the process is reversed. The compressed air is returned to the surface where it is used to burn natural gas in the combustion chambers of the power unit. The resulting combustion gas is then expanded in a gas turbine to spin the generator and produce electricity (Lady & Katz, 1978; Istvan et al., 1982). To date, there are only two CAES plants in the world: a 290-MW plant in Huntorf, in the western part of Germany, built in 1978, and a 110-MW plant in McIntosh, Alabama, USA, commissioned in 1991. Both plants store the compressed air in underground salt caverns (Crotogino et al., 2001). At the time, the main argument for constructing these plants was the requirement of instant load-balancing facilities ('minute reserve', i.e. power-station output that can be made available within a few minutes, as opposed to large conventional power plants that take a few hours to arrive at their full capacity).

Power generation in the Netherlands

The installed electricity-production capacity in the Netherlands currently totals ca. 20 600 MW (2005), of which ca. 16 600 MW are linked to the national power grid. Plans for installing offshore wind farms for a total capacity of 9200 MW have been made public. Due to government subsidies such investments are apparently rather attractive. However, the government is considering to subsidise part of the plans only. Currently ca. 1500 MW onshore wind-energy capacity has been installed (i.e. ca. 9% of the Dutch grid capacity). TenneT, the Dutch power-grid operator, considers as safe a maximum share for wind energy of 10 to 15% of total grid capacity. Above this level, the flexibility of the grid may not be adequate for handling the voltage fluctuations when attempting to balance supply and demand. The fluctuations also present a problem on an hourly basis. In view of the planned increase in the capacity of wind power and its inherently less precise short-term predictability of the required power production, minute reserve will become an essential complement to large-scale wind power. Subsurface CAES could provide this and allow a higher fraction of wind energy to be installed. Conventional CAES, however, has a relatively low thermodynamic efficiency. To improve this, a new type of CAES named 'Advanced Adiabatic CAES', is being investigated (Bullough et al., 2004).

Other forms of underground energy storage

Other forms of underground energy storage have been investigated in the Netherlands. To improve the thermal efficiency of CAES, compressed-air combined with underground heat storage (adiabatic-CAES) was studied (Heederik et al., 1989), but proved uneconomic because of the large number of wells required. Shell investigated 'pumped hydro' using a shallow and a deep salt cavern, but could not resolve the problem of having to install a turbine in a deep subsurface cavern through a narrow borehole. Drilling a shaft to the tight limestone of the Dinantian (Lower Carboniferous) Zeeland Formation has been researched in the province of Limburg with the objective to install a turbine in a subsurface gallery (Rijks Geologische Dienst, 1986). This 'OPAC' facility (Ondergrondse Pomp Accumulatie Centrale) was to be complemented with a water reservoir at surface, but proved not feasible.

Underground solid-waste disposal

The disposal of toxic and radioactive solid waste was investigated in the recent past, but to date no field tests have been carried out. Radioactive-waste disposal in salt diapirs was extensively researched as part of the OPLA project, and through its follow-up CORA project (Visser, 1987; OPLA, 1989; Van den Broek et al., 1996; Ministerie van Economische Zaken, 2001). A historical review of this research and its societal implications was published in Dutch by Faasse (2002). The CORA research focussed on the comparison of three options for retrievable storage or disposal: long-term aboveground storage versus underground storage in either rock-salt formations or deep clay deposits. For each option, the retrievability and safety aspects have been evaluated. The research did not reveal any factor prohibiting the technical feasibility of retrievable disposal. For the underground options, retrievability requires additional facilities. Based on today's knowledge, retrievability can only be guaranteed for a few hundred years. The calculation of worst-case radioactive-pollution scenarios yielded low to very low annual dose rates.

Legal framework

The storage of gas in the subsurface is governed by the Mining Law (Mijnbouwwet) of January 1st, 2003. This act applies to all substances that are injected into the subsurface, whether for temporary or permanent storage. It however does not stipulate the commercial terms and conditions for undertaking storage operations as these are imposed, in rather general terms, by the applicable EU directives. As the energy and gas markets are being liberalised, many commercial details have not yet been clarified. For example, currently confusion exists about the tariff structure for flexibility services such as provided by underground gas storage. The issue 'negotiated vs. regulated Third Party Access' is particularly controversial. The main problem is whether the Dutch regulator DTe (Dienst Uitvoering en Toezicht Energie) will control the tariffs for storage in and supply from the facilities. This will affect the behaviour of investors, and thus the planning of more facilities in the future. With respect to CO_2, the trading of greenhouse-gas emission rights will govern the commercial terms of CO_2 storage. In Europe, this is expected to become effective as of 2010, if the guidelines for monitoring and reporting of CO_2 storage will be accepted by the European Commission. At the current stage, several international pilot-tests are being conducted to gain experience with CO_2-emission trading.

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Geothermal energy

A. Lokhorst &
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ABSTRACT

In the Netherlands, the use of the shallow subsurface (generally down to 500 m depth) for extraction and storage of heat is well established. The use of hot ground water for balneological purposes is modest; two thermal baths use water from Eocene sands at depths between 600 and 750 m, while two others exploit water from Carboniferous limestones at 400 m depth, and from Permian sediments at 900 m depth, respectively. Groundwater at greater depths is a potential source of energy. Based on subsurface temperature data, several evaluation projects were carried out in the 1980s, ultimately resulting in a number of inventory and feasibility studies. As yet, however, these studies did not lead to the actual exploitation of geothermal energy. Aquifers that are of potential interest for heating purposes occur at depths of less than 100 m to more than 3000 m in Permian, Lower Triassic and Lower Cretaceous sandstones and in two Tertiary sand units. In total, ca. 90×10^{18} J (equivalent to 2400×10^9 m^3 natural gas) of heat in place (HIP) may be present in these deep reservoirs. The fraction of this energy that may eventually be produced successfully, however, depends strongly on location-specific reservoir properties.

Keywords: Netherlands, heat in place, aquifers, geothermal gradient

Introduction

The earliest known records of subsurface temperatures in the Netherlands were obtained from a borehole, drilled in 1872 to a depth of 365 m in the city of Utrecht in the centre of the country (Harting, 1879; Visser, 1978). Between 1910 and 1960, temperatures were measured in various coal-exploration wells in the east and south of the country. Temperature data are also available from oil and gas exploration and production wells in the Netherlands onshore and offshore areas. These data show that subsurface temperatures are related to the lithology of the sediments; for instance, the geothermal gradients in the Upper Carboniferous strata tend to be high, over 4 and up to 5.2 and even 5.6°C per 100 m, due to the low thermal conductivity of coal; in contrast, the high conductivities of rock salt and anhydrite cause low gradients of 2.5°C/100 m (Visser & Heederik, 1987).

Van Dalfsen (1981) published country-wide temperature maps for 10 depth levels, from 25 m to 250 m. These maps resulted from accurate measurements of equilibrium temperatures in small-diameter groundwater piezometers. The maps show that natural groundwater flow is effective in perturbing a subsurface temperature field due to pure heat conduction. Additional temperature data allowed an update of the shallow subsurface temperature field (Van Dalfsen, 1982).

Prins (1980) published the first deep-subsurface temperature map of the Netherlands, based on oil and gas exploration data. Updates, based on larger data sets, were prepared by Ramaekers (1991) and Rijkers & Van Doorn (1997), and include maps of various depth levels. Van Balen et al. (2002) presented deep subsurface temperatures in the Roer Valley Graben and Peel Block area, concluding that, contrary to the results of previous mapping, the Roer Valley Graben is probably not a relatively cold area in the Netherlands. The temperature distribution shown in Figure 1 indicates average geothermal gradients of ca. 3 to 4°C/100 m down to 2000 m depth. The average surface temperature is ca. 10°C (Van Dalfsen, 1981).

In the Netherlands, geothermal-energy exploitation from groundwater can be regarded as a potential source of energy. The rise of oil prices in 1973 boosted the interest in this kind of non-conventional energy, resulting in a government-established 'Discussion Group Geothermal Energy' (Visser & Heederik, 1987). This group formulated various evaluation projects, ultimately resulting in a number of inventory and feasibility studies in the 1980s which did not, however, lead to the actual exploitation of geothermal energy. Studies dealing with the temperature distribution, reservoir characterization and required reservoir properties in the shallow subsurface were performed by Groundwater Survey TNO (e.g. Csonka, 1968; Van Dalfsen, 1982; Dufour, 1984). Geological inventory studies for geothermal-energy purposes were carried out by the Geological Survey (RGD, 1982, 1983, 1984, 1985). The geothermal test well Asten-2, drilled in 1987 in the province of Noord-Brabant, was not successful (Visser & Heederik, 1987; Heederik & Huurdeman, 1988). The targeted Tertiary aquifers, the Houthem (1630–1670 m below mean sea level) and Dongen (1500–1550 m) formations, Voort Member (1050–1250 m) and Breda Formation (850–950 m) showed poor reservoir development; the Vessem Member of the Oligocene Rupel Formation at 1490–1510 m below the surface (elevation ca. 25 m above mean sea level) yielded water with a temperature of 54°C, which was considered too low for the heating of greenhouses in the area.

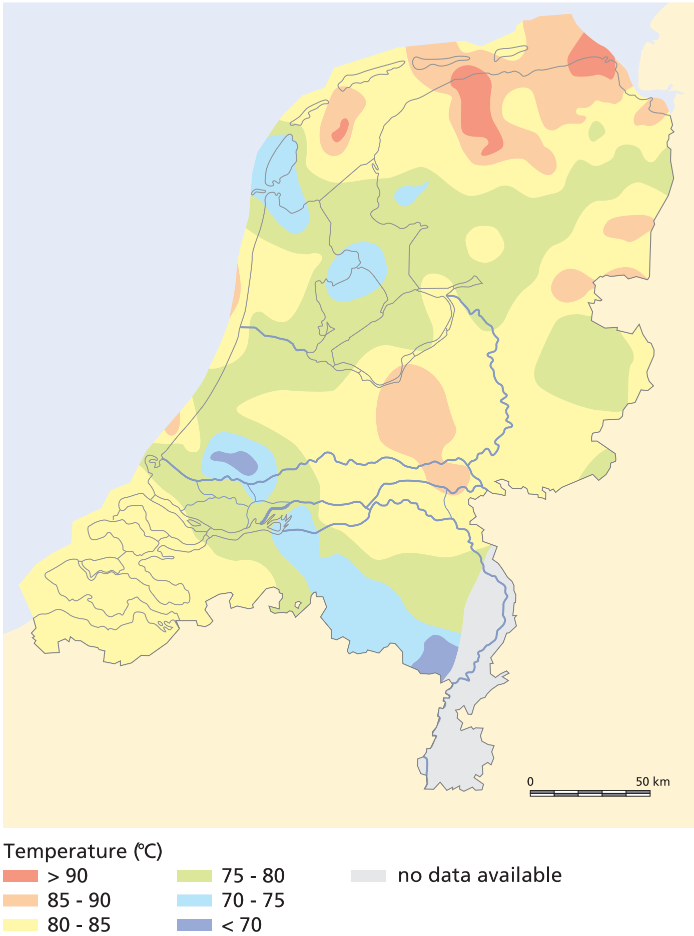

Fig. 1. Temperatures at 2000 m depth below the surface, obtained from measurements in boreholes (NITG, 2004). In the 'no data' area in Limburg, only one well (Nederweert-1) reached a depth of 2000 m.

In the past few years, the potential of geothermal energy as a sustainable energy source received renewed interest, partly as a consequence of the 1997 Kyoto protocol which aims at achieving a worldwide reduction in CO_2 emissions. Moreover, recent oil-price increases have boosted the interest in geothermal energy for application in both district heating and greenhouses.

Geothermal-energy production

From a geological point of view, geothermal energy projects require formation water that is of sufficiently high temperature and a reservoir rock that allows the production of sufficiently large volumes of water. The temperatures required depend on the kind of application. They are, for example somewhat lower for greenhouses than for district-heating systems which generally have an inlet temperature of 70°C. In the Netherlands, water temperatures of 70°C or higher are found in aquifers deeper than 2000 m and temperatures around 45°C in aquifers at depths of 1000 to 1200 m. Figure 1 shows the temperature distribution at 2000 m depth.

The transmissivity, i.e. the product of thickness and permeability of the aquifer, should be sufficient to permit a production of several thousands of cubic metres per day. It is the main risk factor and often the reason that potential aquifers and locations turn out to be unsuitable. In practice, only thick or very permeable aquifers are prospective.

Both in the storage of energy (heat and also cold) at shallow depth and the extraction of heat from greater depth, open or closed systems can be employed (Table 1). The water used in an open system circulates from time to time freely through the aquifer. The liquid used to transport heat in a closed system on the contrary always circulates through various kinds of pipes and vessels.

Shallow applications

In the Netherlands, combined heat and cold storage is applied at many places, using groundwater in shallow aquifers (generally at less than 500 m depth) as the main carrier of the energy in a two-well system. During wintertime, heat is extracted from groundwater produced by a 'hot' well, and after heat extraction above ground, the cooled water is injected through another 'cold' well. In summertime the system is reversed and groundwater is-produced from the 'cold' well, and after cold extraction, the heated groundwater is pumped back through the 'hot' well. In general the total heat and cold added to and subtracted from the aquifer are balanced on a yearly basis. The water temperature in this open circuit ranges from about 5 to 25°C. The thermal power is generally between 200 and 20 000 kW. Most applications are in offices and commercial buildings. More than 300 systems have been installed (Van Heekeren et al., 2005).

A second shallow-subsurface application is a closed system with borehole heat exchangers. In general this is used down to 150 m. A fluid is pumped down a borehole, through the heat exchanger(s), and then back to the surface via a separate tube inside the same borehole. During

Table 1. Geothermal-energy production methods.

Depth range	Open systems	Closed systems
Shallow (down to 500 m)	Heat and cold storage Heat storage	Shallow borehole heat exchangers
Deep (down to > 4000 m)	Low-enthalpy geothermal energy Hot Dry Rock	Deep borehole heat exchangers

wintertime the heat is extracted aboveground from the circulating fluid, gradually cooling the immediate surroundings of the well. In summertime the system is reversed for cooling purposes. Here too a balance between heat and cold is necessary. According to Van Heekeren et al. (2005), over 1100 of this type of systems were in operation by the end of 2004, mainly for small-scale applications such as single-family houses and small-size office or commercial buildings. In the houses, most of the systems are used only for heating purposes. In buildings, however, the systems are mainly applied for both heating and cooling, using the underground both as a heat source and a heat sink. The systems have a thermal power ranging from 50 to 100 kW.

A third shallow application is the storage of heat. Surplus heat is temporarily stored in the subsurface aquifer and is extracted at a later stage. This is realized at a few (< 10) locations only.

Deeper applications

In the applications installed in the deeper subsurface, the extraction of heat is the main purpose. They use geothermal energy in the strict sense. At the moment this kind of application is not realized in the Netherlands. However, the sharp increase in energy prices, combined with Kyoto-protocol measures, leads to an increasing interest in low-enthalpy geothermal energy, and the realization of such a geothermal project is expected within the next few years. The first exploration licence (2006) for this purpose has been granted to a greenhouse farmer in Bleiswijk (Zuid-Holland). High-enthalpy ($T > 180°C$) geothermal energy sources such as geysers and steam fields are not present in the Netherlands.

The application of deep geothermal energy for the production of heat is occasionally (Neustadt-Glewe, Germany, and Altheim, Austria) combined with the production of electricity by means of binary conversion techniques (Organic Rankine Cycles or Kalina techniques). These techniques apply (organic) fluids with low boiling temperatures to drive a turbine. Their efficiency is still relatively low, but the remaining heat is used for heating purposes. Currently the German government promotes this technique by guaranteed high prices for the electricity, boosting the interest in geothermal energy to a high level.

Deep low-enthalpy geothermal systems can be open or closed. Low-enthalpy heat extraction operates by means of a two-well system (doublet), which consists of a production well that taps the water from the aquifer and a second well through which the cooled water is re-injected into the aquifer. Operations involving such a geothermal doublet take place in France, Germany, Austria, Hungary and other European countries. In general, the aquifer has temperatures > 70°C, a relatively good permeability (> 300 mD) and a thickness of preferably > 30 m. Its depth is mostly less than 3400 m. An important, and restricting condition in these operations is the balance between heat supply and heat demand. The drilling of a doublet results in high initial investment costs, which have to be balanced by a sufficiently high demand. A doublet can have a thermal power of up to 10 MW, which is enough for the direct heating of about 4000 houses. Such an amount of power would also be sufficient for one of the larger greenhouses as used in the Netherlands today. The cost of heat transport at the surface calls for a short distance between supply and demand.

A second open system is the Hot Dry Rock (HDR) concept. Several European areas with relatively high geothermal gradients have temperatures up to 200°C at depths of 4000 to 5000 m. Cold water is injected into rocks of low to zero permeability, which have been fractured ('fracked') artificially, allowing cold water to migrate from injection to production well. During its migration the water is heated by the surrounding rock, resulting in production temperatures of 180 to 200°C. The main purpose is to generate electricity by means of binary conversion techniques. At the moment, Hot Dry Rock techniques are experimentally applied in Soultz (Alsace, France). If successful, HDR will have a high potential throughout Europe. A new experimental site is in Gross Schoenebeck (eastern Germany), where a borehole is deepened to 4000 m. The target comprises the volcanics and low-permeability clastic rocks of the Lower Rotliegend. A comparable geological situation occurs locally in the eastern Netherlands as well.

A third application, currently in progress, is the extraction of heat from water in the galleries and shafts of the former coal mines in Zuid-Limburg.

Closed deep systems mainly utilize Deep Borehole Heat Exchangers (DBHE). Such an exchanger extracts heat from rock layers at depths less than 2800 m. Water circulates through a co-axial system, downwards along the borehole casing and upwards via an in-hole production tube. The generated thermal power is about 300 to 500 kW. Operators of this kind of system try to maintain an optimum balance between the production of heat and the influx of heat. Such systems are active in a few places in Germany and Switzerland, where they are used for heating purposes. Here too the distance between the location of the exchanger and the place of heat demand is critical as a large transport distance is costly. The Deep Borehole Heat Exchangers currently in operation are all situated in areas with a normal (30°C/km) geothermal gradient.

Aquifers

Aquifers of sufficient thickness and sufficiently high permeability and temperature, suitable for the extraction of geothermal energy, occur mainly in Noord- and Zuid-Holland, Noord-Brabant, and in the northern and eastern

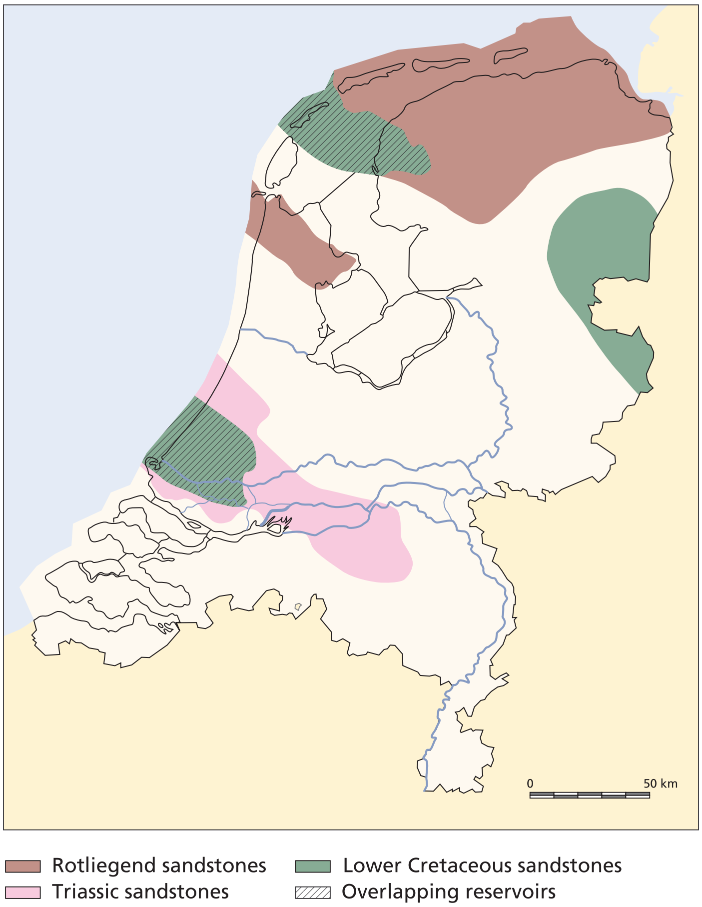

Fig. 2. Distribution of deeper aquifers which are potentially most suitable (T > 60°C and sufficient transmissivity) for the extraction of geothermal energy (cf. Table 2). Localized potential occurrences are not indicated. Neither are Tertiary aquifers which are present under much of the Netherlands (after NITG, 2004).

parts of the Netherlands (Fig. 2). In stratigraphic order, these aquifers are present in:

- Permian, Rotliegend sandstones of the Slochteren Formation in northern Noord-Holland and in Friesland, Drenthe and Groningen;
- Lower Triassic sandstones of the Main Buntsandstein Subgroup in Zuid-Holland and Noord-Brabant, locally also in the eastern Netherlands and Zuid-Limburg;
- Lower Cretaceous sandstones in Zuid-Holland, Friesland and the eastern Netherlands;
- Tertiary sands (Brussels Sand Member and Breda Formation): present at shallow depths under large parts of the country.

The reservoir parameters are listed in Table 2.

Water from the Eocene Brussels Sand Member is locally used for balneological purposes. In 'Aqua Plaza', Ameland island, the sand occurs at 750 m depth and has a temperature of 37°C. In 'Fontana', Nieuweschans (eastern Groningen), its depth is 633 m and its temperature 28°C. The mineral-rich water exploited at 'Thermae 2000' in Valkenburg (Zuid-Limburg) is derived from limestone of the Dinantian Zeeland Formation at approximately 400 m depth; its temperature is 24.5°C (Krings & Langguth, 1987). At Arcen (northern Limburg), mineral-rich water of 42°C is obtained from Permian sediments at 892 m depth. Geothermal energy from aquifers in Pleistocene sands at about 50 m depth is currently applied on a rather modest scale, notably in glasshouse horticulture in Zuid-Holland (Van de Braak et al., 2001).

Potential

The potential for deep geothermal energy in the Netherlands has recently been calculated according to the method proposed by Haenel & Staroste (1988), which determines the 'Heat In Place' (HIP) of the aquifers of interest. Separate calculations were made for the three main intervals, i.e. the Permian Rotliegend, the Lower Triassic and the Lower Cretaceous sandstones. A HIP calculation takes into account the average thickness of a sandstone layer, the average difference between aquifer temperature and surface temperature, and the lateral extent of the reservoir. Moreover, it also involves the heat capacities of the rock matrix and the pore water, which are calculated separately on the basis of the average reservoir porosity.

The results of the calculations, as reported by Lokhorst & Van Montfrans (1988) and Van Doorn & Rijkers (2002), are shown in Table 2. These results carry a considerable degree of uncertainty, in the order of 50 to 80%. They should not, therefore, be used to assess the local potential for geothermal energy; they only indicate an overall figure for the entire aquifer. In total, ca. 90×10^{18} J (equivalent to 2400×10^9 m^3 natural gas) of HIP may be present, i.e. the heat, relative to the average 10°C temperature of the land surface, occurring in the three deeper sand-rich stratigraphic intervals which might be suitable for the extraction of geothermal energy. The amount of this energy that may eventually be produced successfully, however, depends strongly on location-specific reservoir properties.

Finally, it should be noted that the efficiency of transporting heat to users at the surface is limited by the costs of transporting hot water over long distances. The matching of the subsurface-related heat supply and the surface-related heat demand is of utmost importance. The availability of a pipeline infrastructure to transport heat is a major factor in assessing the economic potential of a geothermal energy project.

Benefits and risks

The application of geothermal energy has several benefits, of which the most important are:

- it can contribute significantly towards the reduction of greenhouse gases, because the emission of CO_2 (mainly generated by the necessary pumping) is very low; the emission of other greenhouse gases is negligible;

Table 2. Reservoir parameters and Heat In Place (HIP) for aquifers in sandstone and sand in the onshore Netherlands (sources: Lokhorst & Van Montfrans, 1988; Van Doorn & Rijkers, 2002; NITG, 2004).

Aquifer	Depth (m)	Gross sand thickness (m)	Porosity (%)	Perm. (mD)	Temp. (°C)	HIP (10^{18} J)
Permian, Rotliegend sandstones						
Groningen, Friesland, Drenthe & Noord-Holland	2000–4500	10–200	11–25	30–600	Max. > 100	50
Lower Triassic sandstones						
West Netherlands Basin & Roer Valley Graben (Zuid-Holland & Noord-Brabant)	2000–4000	25–300	Variable	Variable	Max. > 100	30
Lower Saxony Basin (locally)	2000–3500	Max. 80	Variable	Variable	Max. > 100	3
Other areas	300– > 5000	0–50	Variable	Variable	Max. > 100	4
Lower Cretaceous sandstones						
West Netherlands Basin (Zuid-Holland)	700–2500	Max. 250	15–30	Max. 3000	Max. 90	3
Lower Saxony Basin (especially SE Drenthe)	800–1800	3–65	15–20	220–500	40–80	0.4
NW Friesland	1800–2100	10–200	15–22	1–30	70–80	
Tertiary sands						
Brussels Sand Mbr	100–1150	0–135		Max. 600	15–45	
Breda Fm	< 835	Variable	30–35	50– > 200		
Total HIP						> 90.4

– it does not entail visual or noise nuisance;
– the security of supply is high; in principle, a geothermal plant can operate all year long and its capacity is independent of seasonal fluctuations and weather conditions;
– the technology is safe and proven, mainly based on extensive experience in oil and gas production.

A few drawbacks can be foreseen in the Dutch situation. The aquifers, suitable for geothermal exploitation are the same as the oil and gas-bearing reservoirs, implying considerable overlap (Fig. 2). A deep-seated geothermal project positioned close to a producing oil or gas field or a gas or CO_2 storage facility, therefore may cause subsurface interference. The extraction of geothermal energy may affect the pressure distribution in or around the oil or gas field or the storage facility. A recent simulation study (Brouwer et al., 2005) concerning the effects of overpressure and temperature changes in the Lower Cretaceous IJsselmonde Sandstone Member, however, shows that, if re-injection of water takes place under overpressure conditions, the pressure changes in the direct vicinity of the wells due to the extraction of geothermal energy are limited, i.e. no more than 1 bar at a distance of 1 km. The simulation also shows that thermo-elastic effects may occur as well, depending on the temperature of the injected water. These may amount to 50 bar if the formation is cooled by more than 50°C, which would locally cause a compaction of 2 to 3 cm at reservoir level (ca. 1100 m); the effects at surface would be negligible. Such investigations are important in the Netherlands, where public concern about soil subsidence and seismicity due to gas production plays an important role in discussions on the use of the underground.

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