

2. CONSTITUTION OF ROCKS AND ROCK FLUID BEHAVIOUR

2.1 Rocks in general

2.2 Matrix minerals and geochemistry

2.2.2 The presence Elements

2.2.2 The presence of minerals, coal and hydro-carbons

2.3 Geological classification

2.4 Geo-temperatures, geo-Pressures, Time-effects

2.4.1. In-situ temperatures and related salinities

- Geothermal gradient
- Chemical reactions
- Fluid salinity and water resistivity

2.4.2 In-situ pressures

- Pressure gradients
- Overpressures
- Pressure gradients with variable fluid and gas densities.

2.4.3 Effects of time

2.5 Rock bulk densities and matrix densities

2.5.1 Density definitions

2.5.2. Laboratory measurement methods

2.6 Exercises

2.1. ROCKS IN GENERAL

Rock material and rock mass can be classified, from microscopic to scenic scale through its heterogeneity and anisotropy. Conventional concepts of physical behaviour of materials are, with precision, occasionally applicable on rocks. Normally, within the volume of rock being studied and regardless of the scale, the properties vary from point to point (heterogeneity) and with direction (anisotropy). The rock material, or rock substance/intact rock/rock element, is the - to all appearances - continuous substance on which laboratory property tests are run, and to which rigid body continuum mechanics can be practised. The rock material volume element is large regarding grains, pores, and cracks and small when discontinuities like fracs, cleats and layering are considered. The rock mass, or in-situ rock, include a larger mass than does the resembling rock material. In the sense of physics rock mass means mass of rock. However, in nature it consists of heterogeneous material and discontinuities such as pores and fracs. Hence the difference in physical properties between rock material and rock mass justify the distinction between the two.

In general rock consists of minerals, grouped in certain shapes, with openings that can be filled with fluids or gases.

- Minerals are the structure elements of rocks. The structure of the network normally is heterogeneous. Ordinary rock-forming minerals are anisotropic with impurities of other mineral matter. The minerals are connected at grain surfaces, by cements and cohesive aggregates.
- Groups of grains or minerals can be arranged in layers, lenses, etc. and represent in a certain way a part of its geological history, such as grain-size differences due to sediment transport by water or air, or from parallelism of platy grains recrystallized in equilibrium with a stress field.
- At small scale (< 10 cm) rock already is filled with cracks and pores. At larger scales it is cut by discontinuities like fractures (joints) and divisions between layers. The pores and fractures and their orientations are rarely distributed uniformly through rock.
- Pores and cracks can be filled, partly or completely, with water. Water and rock affect each other chemically and physically. Dissolution and precipitation strengthens or weakens rocks. Moreover, a dry rock is a different material from water-saturated rock. In general rock that has been exposed to percolation or weathering normally differs considerably from its unweathered equivalent.
- Although residual stresses are not a part of the composition of rock, stress impact can affect its characteristics. For example, residual stresses affect the amount of load needed to reach failure and may influence the orientation of failure surfaces. Fractured rock behaves differently from its unfractured equivalent. Further fracture properties such as spacing, orientation, size, and roughness are covering the rock heterogeneously. The stronger the rock material, the more important are the discontinuities in determining the behaviour of the rock mass.

In all considerations of the composition of rocks, it is essential to keep in mind the effects of scale. This is particularly important with regard to heterogeneity and anisotropy. Rock material viewed at one scale might be homogeneous and isotropic, but viewed at larger or smaller scales it could be neither homogeneous nor isotropic.

2.2. MATRIX MINERALS AND GEOCHEMISTRY

A small amount of elements form the bulk of the earth's crust. Moreover, an even less number of elements in minerals make up the major parts of rocks. Chemical content and crystal shape determine the properties of minerals and mineral composition and matrix textures define the rock characteristic features. Mineral content and textures are used to explain the bulk physical characteristics of rocks. The crust's mass is about 2.5×10^{22} kg ($\approx 0.4\%$ of the earth mass, Turekian, 1972).

2.2.1. THE PRESENCE OF ELEMENTS

Eight elements make up over 98 wt.% of the oceanic and continental crusts. Even more, Si and O account for 65 % to 75 % of the Earth's crust. The difference in presence of elements projects the variety in differences in geological settings of the crust, to wit:

- More potassium, sodium and silicon in "granitic" continental crust.
- More iron, calcium and magnesium in "basaltic" oceanic crust.

TABLE 2.1: Main elements in the Earth's Crust (from ¹Tan et al., 1970 & ²Tiab et al., 1996)

ELEMENTS in wt.%	Crust			Rocks	
	Ocean ¹	Continental ¹	Earth's crust ²	Igneous & metamorphic ¹	Sedimentary ²
O	45	46	46.40	46	50
Si	23	29	28.15	27	21
Al	8.4	8.3	8.23	8.6	5.6
Fe	7.5	4.8	5.63	6.0	3.3
Ca	7.1	4.1	4.15	4.8	10.0
Na	2.0	2.5	2.36	2.5	0.9
K	1.1	2.0	2.09	1.6	1.9
Mg	3.9	2.1	2.33	2.8	2.0

Table 2. 1 Main elements in the Earth's Crust (from ¹Tan et al., 1970 & ²Tiab et al., 1996)

The volume percentage of elements gives a different view. Oxygen, normally bound with Si and Al, and the largest of the mentioned elements, justifies ≈ 95 vol.% of the earth's crust.

In addition, many elements are dissolved in water (table 2.2.). Chlorine, sodium, sulphur, and calcium are the five most abundant dissolved elements in both sea water and fresh water. In this respect one also can imagine the idea of brine that is stored in pores and vugs in a major part of the sediments and to a large extent in igneous and metamorphic rock. Oxygen is the most abundant element in water and

Main elements dissolved in (mainly from Turekian, 1972)

	Sea water in $\mu\text{g/l}$	Fresh water $\mu\text{g/l}$
Cl	$1.94 * 10^7$	7800
Na	$1.08 * 10^7$	6300
Mg	$1.29 * 10^6$	4100
S	$9.04 * 10^5$	5600
Ca	$4.11 * 10^5$	15000
K	$3.92 * 10^5$	2300
Br	67300	20
C (inorganic)	28000	11500
N	15000	n.a.
Sr	8100	70
O	6000	n.a.
B	4450	10
Si	2900	6100
F	1300	100

by weight almost twice that in rocks.

Table 2. 2: Main elements dissolved in (mainly from Turekian, 1972)

The composition of rocks, water, and air, in and on the crust change in time and are interacting with each other chemically and physically.

2.2.2. THE PRESENCE OF MINERALS, COAL AND HYDRO-CARBONS

A rock forming mineral normally is defined a naturally occurring inorganic crystalline synthesis. According to this definition coal and hydro-carbons are no minerals, but in this course they are included as a division of the rock components. Mineral compositions can be categorised in:

- silicates
- oxides and hydroxides
- sulphides and sulphates
- carbonates
- sulphates
- phosphates
- halides,
- and in our case coal + hydrocarbons.

Silicates as a total are the most present minerals in the earth's crust. In addition, feldspars constitute over half and when quartz is included nearly two third of the total.

<i>Principal minerals in the Earth's Crust (after Ronov and Yaroshevskiy, 1967)</i>	
Mineral	Presence in %, based on the actual mineral composition
Quartz	12
Potash feldspars	12
Plagioclase feldspars	39
Micas	5
Amphiboles	5
Pyroxenes	11
Olivines	3.6
Clay minerals and Chlorite	4.6
Calcite and Aragonite	1.5
Dolomite	0.5
Magnetite and Titanomagnetite	1.5
Other minerals like Garnet, Kyanite, etc.	4.9
Coal and hydro-carbons	accessory
	-----+
Total	100

Table 2. 3: Principal minerals in the Earth's Crust (after Ronov and Yaroshevskiy, 1967)

In order to get acquainted with the most occurring matrix minerals, their composition and densities, they are listed in table 2.4. The mentioned minerals are predominantly present in sediments and to a lesser extent in igneous and metamorphic rocks. Some ore minerals are tabled due to their general presence in many rock categories. However as parts of ore bodies they will not be discussed in this lecture and therefore the major part of these mineral will be neglected in the overview. All tabled minerals are also classified on their occurrences. Igneous and volcanic rocks are formed from solidification of molten crust/mantle substance. Sedimentary rocks are fragmented clastic matter or precipitated chemical substance formed at the surface. Diagenetic/Metamorphic products are developed by the presence of pressure, heat, or both to already existing rocks. In other words, diagenesis/metamorphism is a rock-to-rock transformation.

The matrix minerals form a fabric, which is the sum of all the structural and textural features, including the correlative physical properties, of a group of arranged minerals. Fabric data can be separated into:

- scalar data, which include non-directional characteristics like; grain shapes, grain-size distribution, density, porosity and permeability.
- vectorial data, which are based on characteristics like orientation of grains, fractures/pore shapes, etc. Diagenesis/metamorphism, mineral growth and clastic deposition are mainly responsible for vectorial fabrics.

The mineral composition and fabrics or textures are the major criteria for the geological classification of the various rock types. Fabric analysis is normally used as a classification tool to value rock engineering parameters that are used for, among others, (directional) drilling, bore hole stability, excavations, large (sub-surface) constructions, off-shore installations, etc.

<i>Most relevant minerals, composition, density and main occurrence (from Carmichael, 1985 and Schlumberger 1989)</i>			
Name	Composition	Density kg/m ³ x 10 ³	Occurrence *
Silicates			
Olivine	(Mg,Fe) ₂ SiO ₄	3.2 - 4.14	i, v
Garnet	(Fe,Mg,Ca) ₃ (Fe,Al) ₂ (SiO) ₃	3.75 - 4.25	i, m
Pyroxenes	(Ca,Mg,Fe,Al).[(Al,Si)O ₃] ₂	3.2 - 3.5	i, m
Amphiboles	Ca ₀₋₂ (Mg,Fe,Al) ₅₋₆ [(Al,Si) ₄ O ₁₁] ₂ (OH) ₂	3 - 3.2	i, m
Quartz	SiO ₂	2.65	i, (v), m, s-cl/ch, d
Feldspar	(Na,K,Ca).Al.(Al,Si) ₃ O ₈	257 - 2.76	i, m, v, d, s-cl
Micas	K ₀₋₁ (Mg,Fe,Al) ₃ (Al,Si) ₃ O ₁₀ (OH) ₂	2.7 - 3.2	i, m, (d)
Clay minerals	(K,Na,Ca,Mg) ₀₋₂ [(Al,Si) ₈ O ₂₀](OH) ₂₋₄ .n(H ₂ O)	2.5 - 2.65	s-cl/ch, m, d
Carbonates			
Calcite	Ca.CO ₃	2.72	s-ch/cl, d (i)
Dolomite	(Ca,Mg).CO ₃	2.85	s-ch/cl, d (i)
Siderite	Fe.CO ₃	3.96	s-ch/cl, d
Sulphides & sulphates			
Pyrite	FeS	5.02	i, m, d, s-cl
Galena	PbS	7.6	i, m, d, s-cl
Sphalerite	ZnS	4.1	i, m, d, s-cl
Gypsum	CaSO ₄ .n(H ₂ O)	2.31	d
Anhydrite	CaSO ₄	2.96	s-ch, d
Oxides			
Haematite	Fe ₂ O ₃	5.28	i, m, (v), d, s-ch/cl
Magnetite	Fe ₃ O ₄	5.20	i, m, (v), d, s-cl
(Hydro)-Carbons			
Coal	C:H:O - Antracite; 93:3:4, Bituminous; 82:5:13	1.8 - 1.2	s-cl/ch, d
Oil	n(CH) ₂	0.85	d
Natural Gas	C ₁₋₁ H _{4.2}	0.83 * 10 ⁻³	d
i - igneous; v - volcanic; m - metamorphic; d - diagenetic; s - sedimentary; cl - clastic; ch - chemical			

Table 2. 4: Most relevant minerals, composition, density and main occurrence (from Carmichael, 1985 and Schlumberger 1989)

2.3. GEOLOGICAL CLASSIFICATION

Rocks are normally classified in three main categories:

- igneous and volcanic,
- sedimentary, and,
- diagenetic to metamorphic.

The classification within those groups is based on a mixture of genetic and visual illustrative factors, e.g. mineral composition and texture in a hand specimen. Based on the mineral composition an chemical composition can be estimated and rocks can be classified. However minerals are grown in an

environment with specific physical and chemical conditions, thus knowledge of the mineral composition affords information on the history of the rock and its constituents; especially igneous rocks. If minerals are difficult to determine physical appearances like colour and lamination are used. (e.g. pink granite, black shales, red sands, etc.). Textures normally are connected with particle size, shape and arrangement. The three categories, as mentioned above are related as shown in figure 2.1.

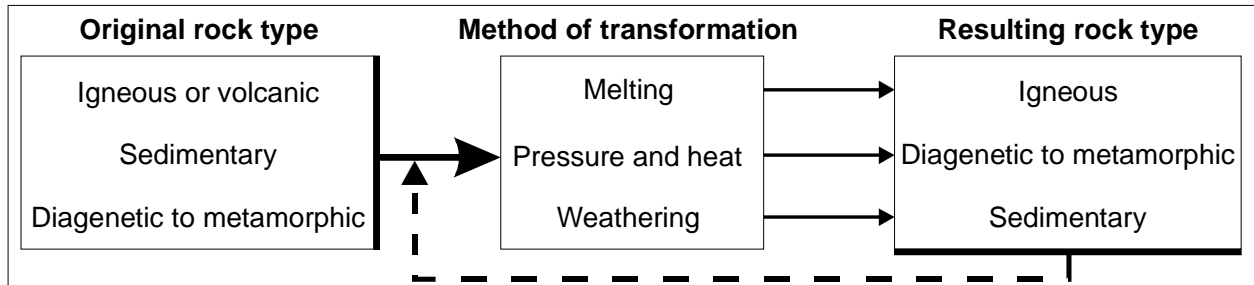


Figure 2. 1: The generic relationship of rocks (revised after Ho et al., 1989)

In this course the emphasis is placed on sedimentary rock. Mining engineers, engineering geologists and petroleum engineers are mainly interested in the physical and geological interpretations. Geologic classification and mapping greatly enhance the reliability of interpolated property data used in engineering classifications. Moreover, a major part of their engineering activities are related to sediments or diagenetically altered sediments. In the coming tables 2.5 a, b and c, for normal sedimentary rocks principal mineral constituents, clastic grain sizes, chemical variety and (bio) organic particles are classified and grouped. Here the most striking particle feature is used to classify the rock. Often sediments are classified in more than one of the main categories, e.g. Oolites, cemented sandstone, fossil bearing limestones.

<i>A general classification of sedimentary rocks, based on composition and texture.</i>		
<i>A: Classification based on clastic fragments</i>		
Texture; grain size	Group name	Composition
Cobbles, pebbles, granules	Blocks to gravel	Rounded & angular fragments
Coarse; > 2 mm	Breccia	Large angular fragments
	Conglomerate	Large rounded fragments
Medium; 2-0.0265 mm	Sandstone	Quartz, feldspar, clay Quartz, feldspar, micas Quartz (feldspar)
Fine; 0.0265-0.0039 mm	Silt	Mainly quartz and clay Quartz, carbonate cement
Very fine; < 0.0039 mm	Clay	Compacted clay Clay, organic matter and some sulphides

Table 2. 5: A general classification of sedimentary rocks, based on composition and texture.

Note that table 2.5 can be regarded as an introduction to rock classification systems and not as a complete reference.

Besides the geological rock classification systems the potential engineering use of the rock or of space in the rock is an engineering classification of rocks. These classifications normally apply to rock material and to rock mass. The related rock properties or material parameters are measured, in a usual way, in a laboratory, and called index properties. Some examples are; porosity, permeability, unconfined compressive strength and specific gravity. The characteristics of a rock mass determined in situ often has an identical name as the laboratory-measured property of rock material. Be aware that the tow measurements cannot be regarded as equivalent measurement results. Compilation of data and application of classification criteria must take into account the way in which each data point has been obtained. Depending upon the needs of the user, classification may be based on one criterion or on a

combination of several criteria. The majority of subjects in this course, with respect to rock measurements, are based on these viewpoint.

2.4. GEO-TEMPERATURES, GEO-PRESSURES, TIME-EFFECTS

B: Classification based on chemical components		
Group name	Example	Composition
Limestone	Chalk	Micro fossils, micro parts of shells etc., clay
	Boundstone	Fossil components bound together by organic growth
	Crystalline limestone	Crystal grains of calcite with relics of ghost structures.
Dolomite	Dolomitic limestone	Dolomite
Chert	Chert	Micro-grains of hydrated silica, light coloured
Flint	Flintstone	Micro-grains of hydrated silica, dark coloured
Evaporites	Gypsum	Gypsum
	Rock salt	Halite, sylvite, aragonite
C: Classification based on bio-organic components		
Group name	Example	Composition
Coal	Peat	Decaying wood embedded in disintegrated plant debris
	Brown coal or Lignite,	Humic (<40 % water) coal with vegetal structures
	Bituminous to Antracite	Organic sediment (< 40% ash) composed of polymers of cyclic hydrocarbons and low water content
Asphalt	Tar	Migrated and solidified immature petroleum. Black to dark brown.
Limestone	Coquina	Fossil fragments, loosely cemented like sand.

Table 2.5.continued: Classification based on chemical components (B) and on bio-organic components (C)

Many sub-surface activities in rocks and its pore space are depending on:

- In-situ temperatures, which normally depend on the geothermal gradient and in that relation the rock type and burial history.
- In-situ pressures, which depend directly on the thickness and density of the overlying rock. Indirectly the stress is depending on the tectonic history, e.g. regional compression zones, shear zones, extension zones and local fault and folding activities. In relation in-situ pore pressures are depending on the height and density of the fluid column. In addition, the presence of sealing layers can develop areas with over pressures.
- Time. All chemical and physical activities in rock need time to proceed. Some effects, like compaction and pressure solution are relatively long term phenomena. Dissolution of carbonates and precipitation of cement or clays could be short term phenomena.

2.4.1. IN-SITU TEMPERATURES AND RELATED SALINITIES

Many processes, related to the in-situ temperature, are constantly changing the rock characteristics through time. Three thermal relevant processes of major importance will be discussed.

Geothermal gradient

Geo-temperatures in-situ depend on the geothermal gradient (G_t) and in that relation the rock type and burial history. In western Europe, in the predominantly Palaeozoic and younger rocks, G_t is about 30°C/km. In shields of Cambrian and older rocks, G_t can be much lower, i.e. 10°C/km. In a crust of approximately 35 km thickness the temperature can reach to some 1000 °C. In the oceanic crusts G_t ranges from circa 40°C/km at the margins to > 200°C/km at the spreading axis or other active plate margins. The normal thermal gradient measured all over the world has an average of $G_t = 18.2^\circ\text{C}/\text{km}$. In this course sedimentary basins are the main type of rock in our topics. If from a certain area the

mean surface temperature; T_s , the formation temperature; T_f (boreholes, mines) and depth; D , are known, then one easily estimates the geothermal gradient (G_t):

$$G_t = \frac{T_f - T_s}{D} \quad (\text{eq. 2.1})$$

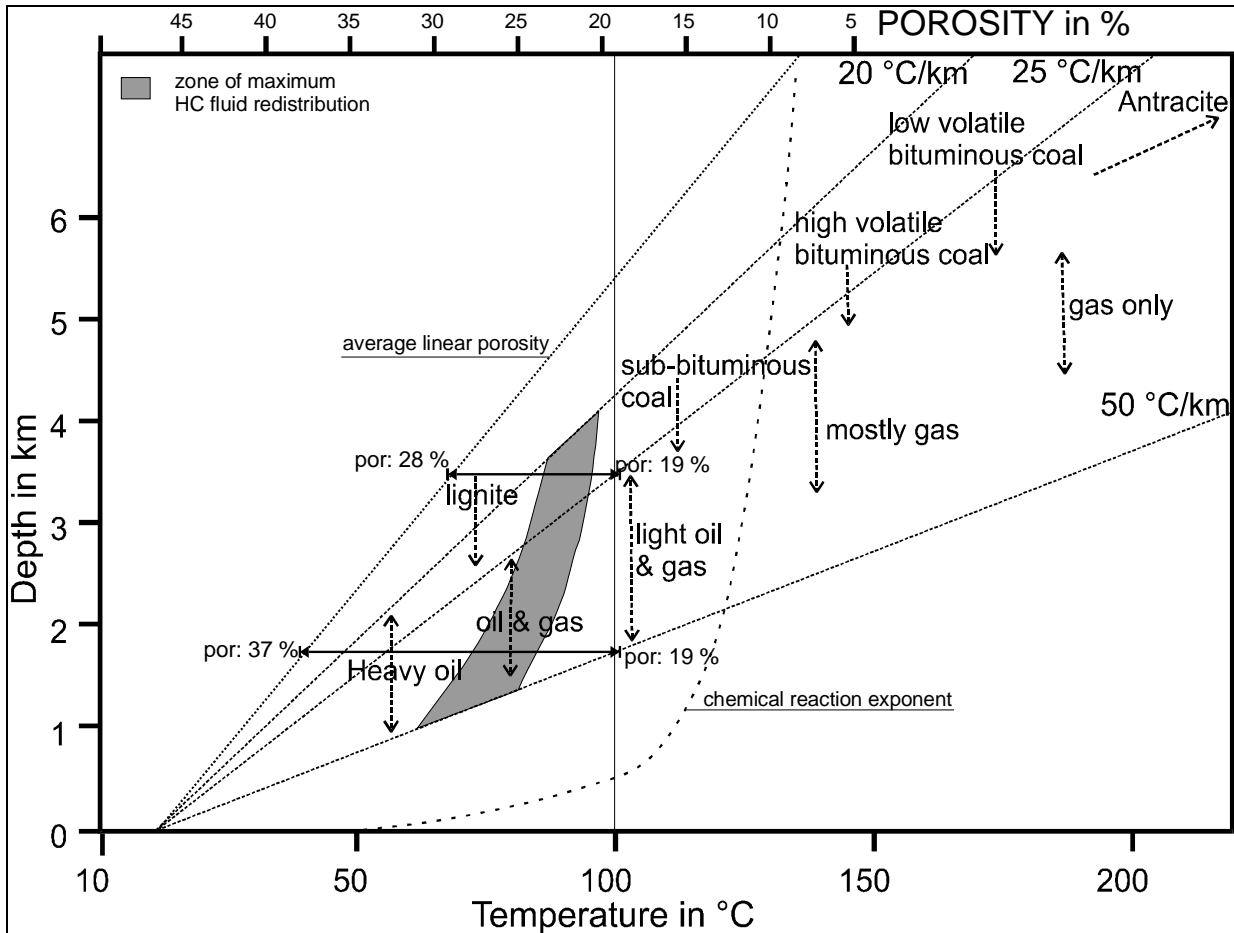


Figure 2. 2 Relation between the depth, temperature, formation of coal, hydrocarbons and the development of porosity.

Chemical reactions

In figure 2.1. the effects of temperatures are related to the maturity of coal and the generation of hydro-carbons. Important are the thermal behaviour of fluids like water and the chemical reaction of minerals and organic matter with or without fluids. These chemical reactions often are depending on temperatures, which can be expressed in the Arrhenius equation:

$$k = Ae^{-\left(\frac{E_a}{RT}\right)} \quad \text{or} \quad \ln k = \ln A - \frac{E_a}{RT} \quad (\text{eq.2.2})$$

where:

- k : the rate constant
- A : pre-exponential factor, depending on environment and mineral types
- E_a : the activation energy of the reaction (J)
- R : gas constant (J/K)
- T : Temperature (K)

The equation shows that the rate of reaction more or less doubles with every 10 °C in rise of temperature.

Fluid salinity and water resistivity

For water, normally present as a brine and sometimes as fresh water (table 2.2), the temperature,

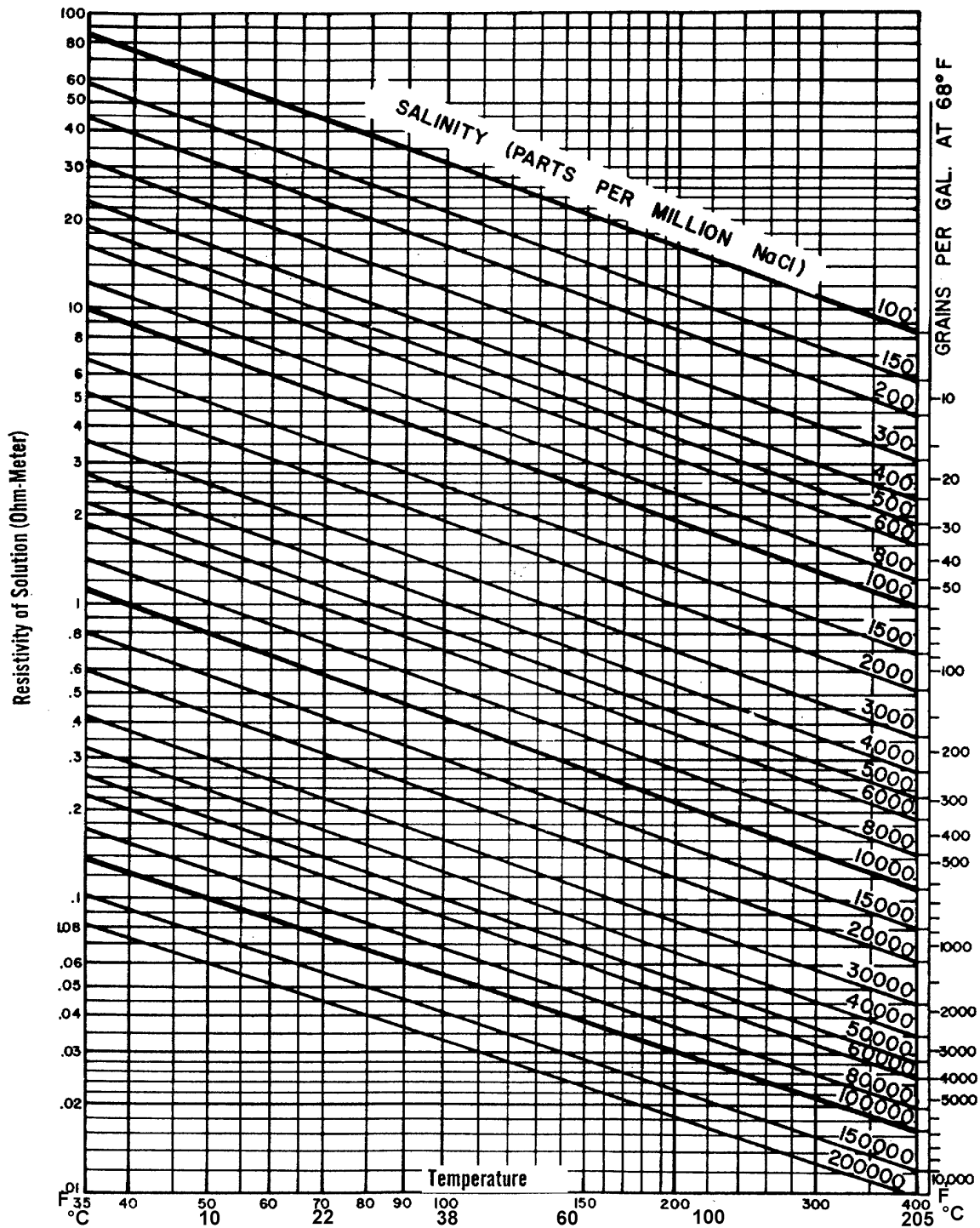


Figure 2. 3: Water resistivity as a function of temperature and salinity (from the Schlumberger log interpretation chart books)

salinity and related water resistivity are important. They have to be determined for environmental corrections in bore hole measurements, for example, when the water salinity at surface conditions has

to be estimated. The resistivity (ohm.m) or conductivity (mmho)¹ of water; R_w can be estimated on the basis of fluid salinity. The solutions are expressed in ppm NaCl-equivalents. Parts per millions (or ppm) is the ratio of weights, e.g. expressed as g/l or $\mu\text{g/g}$, and related as:

$$S_1 = \frac{S_2}{\rho_{sol}} * 1000 \quad (\text{eq.2.3})$$

with: S_1 : salinity in ppm, S_2 : salinity in g/l and ρ_{sol} : solution density in g/l. Figure 2.3 shows the graphic relation between the dependency of R_w from salinity and temperature. It shows that water resistivities decrease with increasing temperature. Here Arp's empirical relation gives for R_w an estimation of the temperature dependency:

$$R_{wT2} = R_{wT1} \left(\frac{T_1 + 6.77}{T_2 + 6.77} \right) \quad (\text{eq.2.4})$$

where, $T_{1,2}$ are the temperatures at specific depths (in °F) and $R_{wT1,2}$ (in ohm.m) the respective water resistivities. For saline water a constant (or correction factor) of 6.77 is inserted. This correction is eliminated when fresh water is dealt with. If the surface temperatures are taken as a standard, T_1 will be 24°C or 75°F and any in-situ R_w -value is estimated for its salinity. Now equation 2.4 can be generalised as:

$$R_{wT2} = R_{w75} \left(\frac{81.77}{T_2 + 6.77} \right) \quad (\text{eq.2.5})$$

Normally a brine consists of a complex sum of ions in a solution. In the previous text we mentioned NaCl-concentration equivalents, or C_{sum} . This concentration is the summation of all ions (n) present

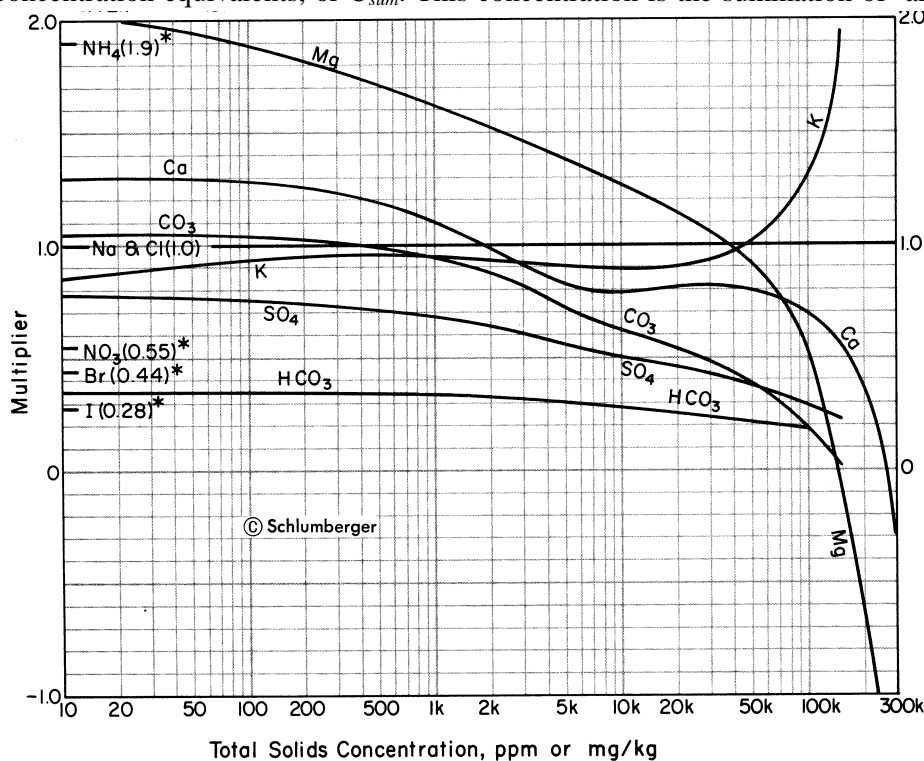


Figure 2. 4: Multipliers for dissolved salt components to equivalents of NaCl concentrations (from the Schlumberger log interpretation chart books)

in the solution as different ions (C_{ai}), multiplied by their weighting multiplier (M_a), or:

¹ 1 mmho = (ohm.m)⁻¹ x 10³

$$C_{sum} = \sum_{a=1}^n M_a C_{ai} \quad (\text{eq. 2.6})$$

The multipliers are determined experimentally and gathered in figure 2.4. When ion concentrations are known from chemical analysis, it is also possible to estimate the R_w at surface condition (24°C or 75°F) with the surface temperature specific relation:

$$R_{w75} = \left(2.74 \times 10^{-4} C_{sum}\right)^{-1} + 0.0123 \quad (\text{eq. 2.7})$$

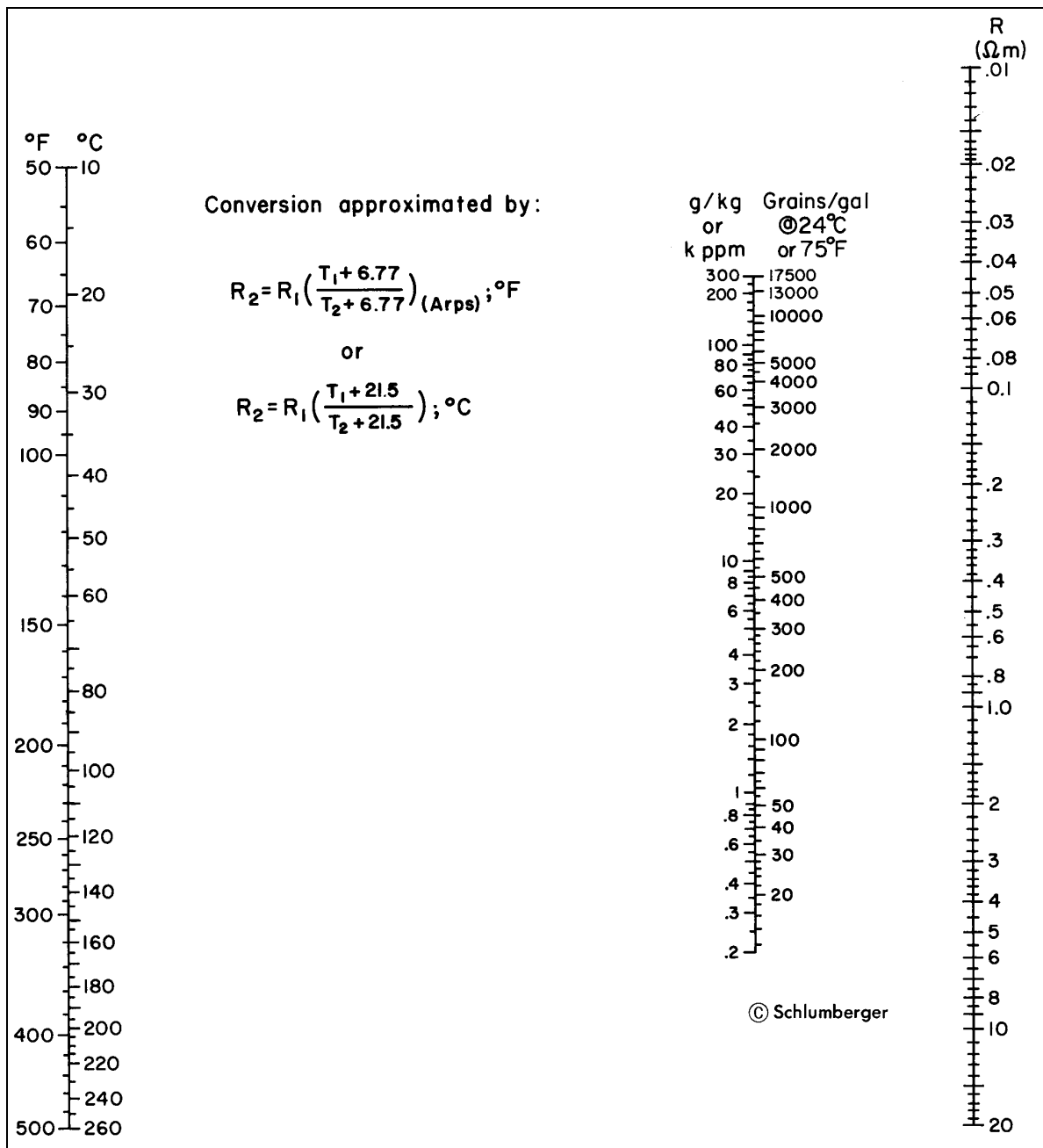


Figure 2. 5: Resistivity nomograph for NaCl solutions. From the Schlumberger log interpretation chart books.

Figure 2.5 gives a second and faster method to relate salinity in NaCl-equivalents with temperature and resistivity. This graphical method shows that each water resistivity at a specific temperature ($T_{1,2,\dots,n}$) can be estimated when a straight line is moved from the temperature line through a point of

constant concentration on the salinity line. This method makes it easy to convert surface conditions to in-situ conditions.

2.4.2. IN-SITU PRESSURES

In-situ pressures, directly depend on the thickness and density of the overlying rock. Indirectly the tectonic history also contributes to anomalies in the stress-distribution. Further the omnipresent pore pressures are related to the height and density of the fluid column. In addition, the presence of sealing layers may develop zones with overpressures.

Pressure gradients

The total overburden pressure (P_o) or confining stress in a rock under compaction, at a specified depth, can be expressed as the sum of the lithostatic (grain) pressure (P_r) and the fluid pressure (P_f). In other words, the lithostatic weight is supported by the pore fluids and its potential energy is stored in the fluids.

$$P_o = P_r + P_f \quad (\text{eq.2.8})$$

In homogeneous systems pressures increase linear with depth. The lithostatic pressure gradient depends on the rock densities, whereas the geostatic pressure gradient depends on the densities of the rock matrix and pore fluids. Hence, the grain framework provides the effective stress (σ_p) in a rock, which can be expressed as:

$$\sigma_p = P_o - P_f \quad (\text{eq. 2.9})$$

In theory the effective stress (σ_p) is equal to the matrix stress (σ_m) or load on the grain framework. The average overburden pressure gradient (rock) is approximated at 22.6 kPa.m^{-1} , which corresponds

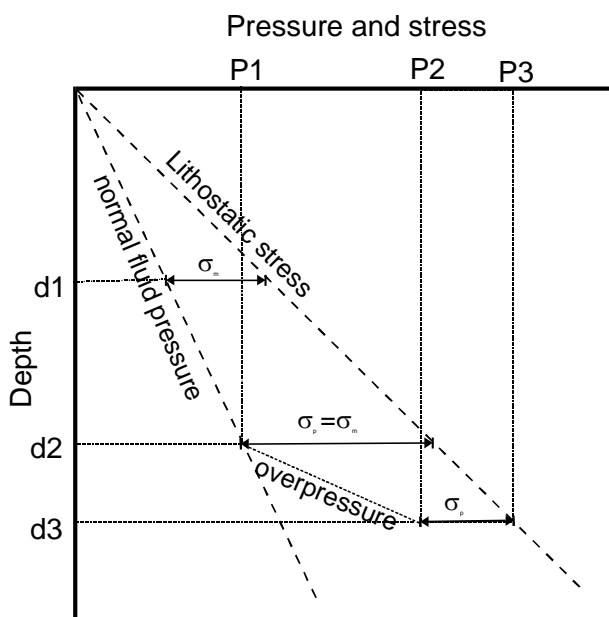


Figure 2. 6: Pressure -depth relation for normal compaction and non-equilibrium.

with an average bulk density of $2.31 \cdot 10^3 \text{ kg.m}^{-3}$. A common fluid pressure gradient is in between 9.8 kPa.m^{-1} (fresh water) and 10.5 kPa.m^{-1} (brine). Normal fluid pressures have the ability to equalise pore fluids in rock with the related hydrostatic pressures for a specific depth. Overpressures have a gradient higher than 10.5 kPa.m^{-1} .

Overpressures

Overpressures (figure 2.6) are not constrained to depth or age. They need an implicit dynamic fluid system (i.e. porous rocks) enclosed by a sealing rock type. Most overpressured zones are present in young basins which are rapidly filled in a regressive regime. For example; near shore sediments are fast accumulating on off-shore mud rocks. Now the increasing fluid pressure gradient forces fluids to upward or side away.

Decreasing porosity and permeability prevent the fluid to migrate and pore pressure built up

starts. The driving force is the load of the water filled overburden.

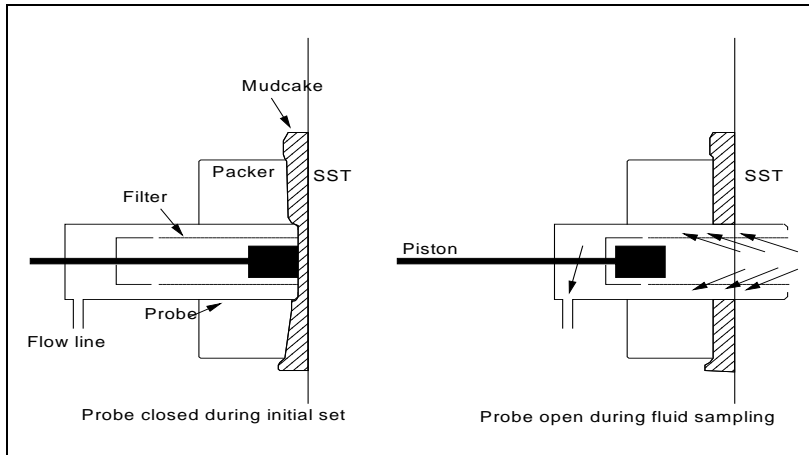
Several types of geopressures can be named:

- Load pressure due to rapid increase in overburden, without fluid escape.
- Tectonic pressure, by thrust or gravity movements of sediment zones, or folding.
- Inflated pressure, induced by fluid recharge

- Phase pressure, due to the change of matrix minor phases or pore fluids like hydro-carbons.

Pressure gradients with variable fluid and gas densities.

In-situ borehole measurements (Wireline Logging, Formation Testing) give a direct indication of the in-situ pore pressures ($P_{f1,2,3,\dots}$) and as a result indirect indications on the fluid types that occupy the pore space. Some decades ago a repeat formation tester (RFT) was developed, which consists of a sample chamber and a pad (figure 2.7). The pad is pressed against the borehole wall and a perforating charge is used to get communication between the porous rock and the sample chamber through a flow line.



In a second option a nozzle is extended through the mudcake into the rock. Now it is possible to measure the formation pore pressure and to get samples from an open hole. The flow-line that connects the nozzle or probe to the chamber contains a number of valves which give way to multiple settings of the tool and collection of several formation fluid samples. At each setting the pressure of the reservoir fluid can be measured with a strain-gauge or a quartz high resolution pressure gauge. The accuracy of the setting of the test tools at a certain depth is controlled by other borehole measurements tools (i.e. Gamma Ray and SP). The reservoir pressure measurements plotted against the depth gives a gradient that is very steep for gas, much flatter for oil, and a small angle for water. The pressure measurements are used to detect contacts such as fresh water/brine, gas/water, gas/oil or oil/water contacts (fig. 2.8).

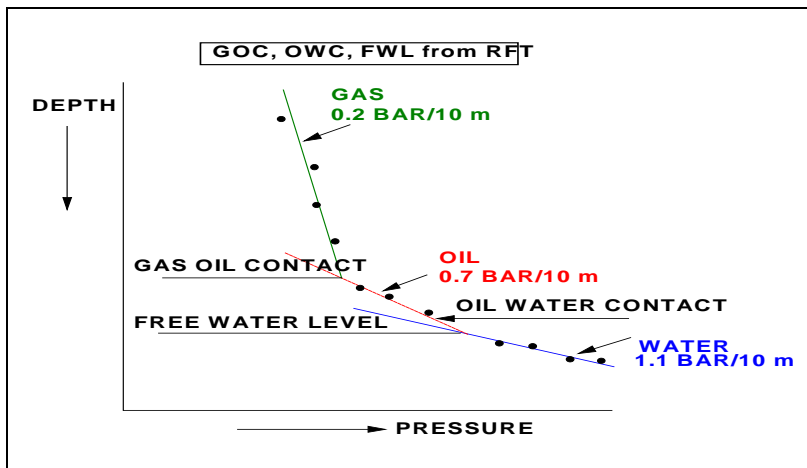


Figure 2. 8: Fluid contacts that can be obtained with fluid pressure measurements.

2.4.3. EFFECTS OF TIME

Time. All chemical and physical activities in rock need time to proceed. Some effects, like compaction and pressure solution are relatively long term phenomena. Dissolution of carbonates and precipitation of cement or clays could be short term phenomena. The relation between temperature and time is essential for the formation/recrystallization of minerals, the maturation of oil and the coalification of plant relics. A simple rock maturity relation, originally by Vassoyevich, included pressure (P in atm) , time (t in million years) and temperature (T in °C). This “geochronothermobar” (G) is defined as:

$$G = \frac{T \times t \times P}{1000} \tag{eq. 2.10}$$

Lopatin (1971) developed a time-temperature index (TTI) for the maturity of organic matter. He covered a depth-age plot with a simplified temperature grid. Here the increment in maturity (ΔM), for a specific depth interval, is related to the exponential relation:

$$\Delta M = \Delta T r^n \quad (\text{eq. 2.11})$$

Here n is an index value for temperature intervals of 10°C . For $100 - 110^\circ\text{C}$ the value $n = 0$ was chosen. Therefore, lower intervals had a $n < 0$ and higher intervals a $n > 0$. The temperature factor r^n consists of the value r , which is based on the type of organic matter and the previously explained exponent n . If the Arrhenius equation is applied (eq. 2.2), with the conclusion that the rate of reaction

doubles with every 10°C rise in temperature, then $r = 2$. The time span (ΔT) is reconstructed from isotope findings or palaeontology.

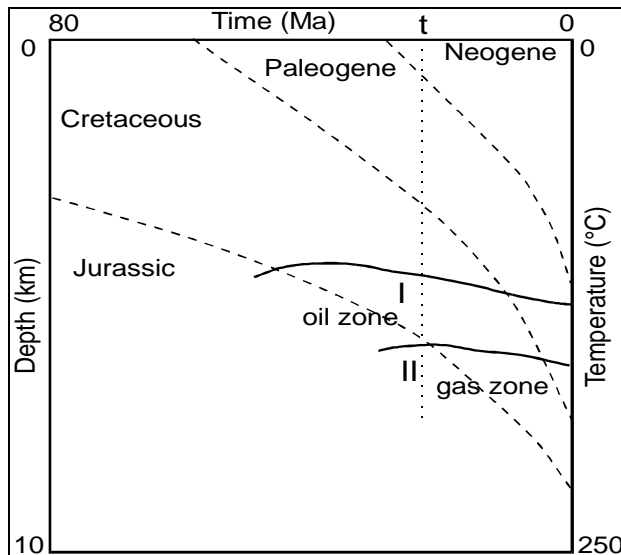


Figure 2.9: Example of the combined role of time and temperature in the maturation of organic matter. Revised after Lopatin, *Int. Geol. Review*, 1980)

Figure 2.9 shows, in an example, that porous rocks of “age I” have been in the oil zone for a period of t Ma. In the same time span porous rocks of “age II” have been in the gas zone.

2.5. ROCK BULK DENSITIES AND MATRIX DENSITIES

With the exception of bulk density most of the aspects regarding mineral densities, textures, fluid densities and related in-situ environments have been discussed in this chapter. Density is the specific rock property that represent the coherence between mass and unit volume. It is used as an index property or as an independent variable to predict other rock properties. However, uniformity of density values are

difficult to obtain, because of the environmental effects, such as; temperature, pressure, kind of fluid and degree of fluid saturation and mineralogical alterations. Further different methods of density determination can give a variation in density values for what seemed to be similar rock types.

2.5.1. DENSITY DEFINITIONS

The American Society of Testing and Materials Standards provided various descriptions for density and specific gravity. In these formulations the temperatures of the material and fluid that is used in the test has to be specified.

- **True density or matrix density:** The mass of a unit volume of a solid material where the volume of only the impermeable portion is considered, represented as:
$$\frac{\text{mass}}{V_{\text{total}} - V_{\text{pore}}}$$

(eq.2.12)

- **True Specific Gravity:** The ratio of the mass of a unit volume of solid material, of the impermeable part, to the mass of the same volume of gas-free distilled water.
- **Apparent Density:** The weight in air of the non-permeable part of the unit volume of rock.
- **Apparent Specific Gravity:** The ratio of the weight in air of a unit volume of material to the weight in air of equal density of an equal volume of gas-free distilled water. Represented as: (dry weight in air)/(dry weight in air - submerged weight).

- **Bulk Density:** Often considered to be equivalent with the apparent density. The weight in air of a unit volume of a permeable material including both permeable and impermeable voids normal to the material. This is expressed in two ways:
 1. (weight)/(volume including all voids) and
 2. (weight)/(unit volume, including the fluid).
- **Bulk Specific Gravity:** The ratio of the weight in air of a unit volume of a permeable material (including all voids) to the weight in air of equal density of an equal volume of gas free distilled water. Here represented as:(dry weight in air)/(saturated weight-submerged weight).
- **Grain Density or matrix density:** Often considered to be equivalent with the true density. Defined as the mass of a unit volume of grains. Also represented as (mass of grains)/(volume of grains) or represented by the formula:
$$\rho_g = \sum_{i=1}^{i=n} \rho_i \cdot v_i \quad (\text{eq.2.13})$$

n is the number of minerals, ρ_i the grain density of each mineral, v_i the volume of each mineral.

Measurement of grain densities decrease the volume effects of pores and fractures. Frequently from density data it is not known if their "true," "apparent," or "bulk" measurements. If the measurement units (g/cm^3) are not presented than the data presumably represent specific gravity (a ratio). If contrarily, then density is intended to be used. Generally measuring methods that are put to use also deal with a measurement of the pore volume, the bulk volume and in consequence the apparent grain volume (both grains and isolated pores). Note that the distinction between bulk volume and apparent grain volume is equal to the volume of interconnecting pores.

2.5.2. LABORATORY MEASUREMENT METHODS

Bulk densities and matrix densities are calculated using the bulk weight or matrix/grain weight and the related volume.

In a laboratory one can measure three different bulk densities:

1. Dry bulk density:
$$\rho_b = \frac{W_g}{V_b}, \quad (\text{eq.2.14})$$

in which W_g is the weight of grains or matrix and V_b is the volume of grains and pore space.

2. Natural bulk density:
$$\rho = \frac{W_g + W_w}{V_b}, \quad (\text{eq.2.15})$$

where W_w is the weight of the available pore fluid.

3. Saturated bulk density:
$$\rho_s = \frac{W_g + (V_p \cdot \rho_w)}{V_b}, \quad (\text{eq.2.16})$$

with V_p as the volume of interconnecting pores and ρ_w as the fluid density.

The bulk weight of matrix and pores or grains is easily and accurately measured with a balance. The procedures used to get a matrix density or bulk density involve the determination of any two of either pore volume, bulk volume, or grain volume.

Bulk volumes can be measured with several methods:

- Volumetric dimensioning of symmetrically shaped specimens (cubes or cylinders) with a ruler or sliding gauge.
- Direct measurement of liquid displacement by submersion of the sample in a non-penetrating fluid; normally mercury.
- Direct measurement of liquid displacement by submersion of a waxed or impermeable sample in a fluid.

- The buoyancy method, based on Archimedes' principle. It requires the weight of a sample in air, both dry (W_1) and saturated (W_2), and its weight hanging in a liquid (W_3). The liquid is of known density (ρ_w). Here the dry bulk density is calculated as:
$$\rho_b = \frac{W_1}{W_2 - W_3} \cdot \rho_w \quad (\text{eq.2.17})$$

and the grain density as:
$$\rho_g = \frac{W_1}{W_1 - W_3} \cdot \rho_w \quad (\text{eq.2.18})$$

Methods for the determination of grain volumes are:

- Fluid displacement of submerged pulverised grains. The pulverised minerals are disposed of major part of the pores and fracs.
- The buoyancy method. The method uses the weight of grains and weight of the specimen that is suspended in a liquid of known density.
- Boyle's law; gas volumetry methods; figure 2.10. Several kinds of porosity-meters, using the pressure-volume relationships of Boyle's law for perfect gases, have been invented. The devices, like the pycnometer, use air or a non-adsorbing gas to get the highest accuracy in analysis. Helium, the best option, invades all void spaces except those that are impermeous. Hence, its results are representative for matrix and grain volumes and they provide the total porosity.

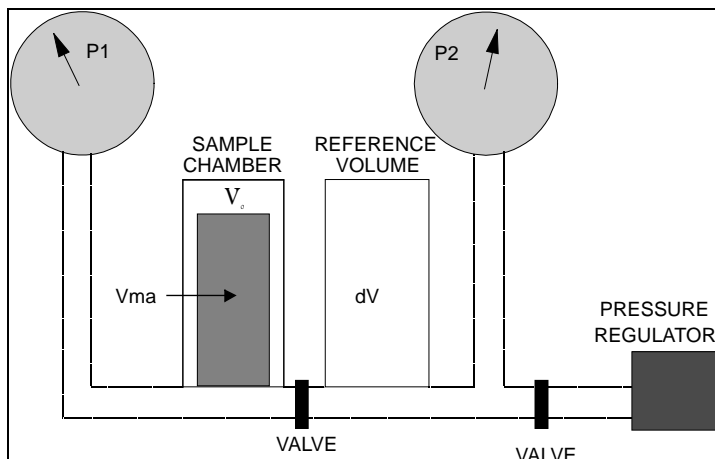


Figure 2. 10: Boyle's law porosimeter, grain-volume determination.

use Boyle's law and use:

$$(V_0 - V_{ma}) \cdot P_1 = (V_0 - V_{ma} + dV) \cdot P_2 \quad (\text{eq.2.19})$$

Now V_{ma} can be calculated, though be aware that isolated pores, common in carbonate rocks, will not be accounted for.

The working method is as follows:
A cleaned and dried sample is placed in a sample holder with a gas pressure P_1 . When the valve that connects the sample chamber with the reference volume chamber is opened the gas will expand isothermally and equilibrate on a pressure P_2 . The volume of the sample chamber is V_0 , and that the core sample has a grain volume V_{ma} . For that reason, the sample chamber will contain a gas volume of $V - V_{ma}$. If this gas pressure changes from the pressure P_1 to P_2 by adding the reference volume dV (normally the same as V_0) we can

2.6. EXERCISES

Bad Bentheim, accommodates a famous German "Kuhrtort" for patients with serious skin diseases. Unfortunately the medical springs are depleting. Nevertheless, newly drilled exploration wells show highly porous sand bodies and fractured carbonate horizons in Jurassic sands and Buntsandstein formations, thus new prospects. The Delft Engineering Geological Group is asked to give an opinion on these findings, with respect to salinity, conductivity and reservoir quality of the drilled wells.

1. For the determination of the water salinities of the different sands, the following salinity results have to be converted to water resistivities at the formation depths e.g. formation temperatures. All resistivities are in ohm.m, FT is the formation temperature in °F or °C, 1 mol Na = 23 gram, 1 mol Cl = 35.5 gram). Use table Ex.1.
2. At the faculty all relevant information on borehole depth and bottom hole pressures have been stored in one computer. Unfortunately, during renovations this specific computer was stolen. Someone found in the literature the regional thermal gradients, with an average estimated gradient of $G_r=30.5^\circ\text{C}/\text{km}$. The average surface temperature is 14°C . Calculate the bottom hole depths and put them in table E1.
3. Now it is also good to know the bottom hole pressures for engineering purposes. Here the salinity and depth are needed.
4. In order to get acquainted with the stability of the porous sandstones, one likes to know the effective stress σ_p on the grain framework. An average bulk density (σ_r) of $2.31 \text{ g}/\text{cm}^3$ is normal in these areas.

Well nr	Bottom hole depth in km	Formation water salinity	ρ_f in g/cm^3	P_f in bar	P_0 in bar	ft	Rw at ft in ohm.m	σ_p in MPa
1		200,000 mg/l NaCl				212 °F		
2		100,000 ppm NaCl				122 °F		
3		50,000 ppm NaCl				130 °F		
4		50,000 mg/l Cl				122 °F		
5		100 g/l NaCl				100 °C		

5. A sixth well was analyzed on the ion-content of the formation water at a surface temperature of 75°F . Further information was vanished. Further information was vanished. The concentrations are: 8.000 ppm Na^+ , 10.000 ppm Cl^- , 5.000 ppm Mg^{2+} , 3.000 ppm Ca^{2+} , 7.000 ppm CO_3^{2-}
 - Determine the solid concentration and the equivalent NaCl concentration.
 - Determine the resistivity at reference temperature and resistivity of the brine at 130°F .