Chapter 9

Heat Transport in Sedimentary Basins

Knut Bjørlykke

The temperature increases downwards in the crust and there is therefore a transport of heat upwards, referred to as the heat flow. Most of the flow is by conduction (thermal diffusion). Flow of porewater will also transport heat in the subsurface but the flow rates in sedimentary basins are normally so small that we can ignore the contribution from fluid flow. Around igneous intrusions there is usually thermal convection with high flow rates and heat transport. In shallow areas with high flow rates of meteoric water, advective heat transport is also significant.

The source of the heat is mainly radioactive processes which are particularly important in the continental crust due to the enrichment of uranium, thorium and potassium in granitic rocks.

Heat is transported through sedimentary basins mostly by conduction following the heat flow equation:

$$Q = c \times {}^{\mathrm{d}T}/_{\mathrm{d}\,z} \tag{9.1}$$

The thermal conductivity (c) is expressed as $\mathrm{Wm^{-1}{}^{\circ}C^{-1}}$ or $\mathrm{Wm^{-1}{}^{\circ}K^{-1}}$ which is the heat (W) transported over a given distance (m) with a certain drop in temperature (°C). Temperature is expressed as Celsius (°C) or Kelvin (K) but Fahrenheit was and still is commonly used in the USA. Conductivity may also be expressed in terms of calories (cal), which is an alternative unit for energy/heat. 1 W equals 1 J/s or 0.239 cal/s.

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The heat flow (Q) is most commonly expressed by $\mathrm{W/m^2}$ but can also be expressed as $\mathrm{cal/cm^2s}$ or $\mathrm{joule/cm^2s}$. A heat flow of 70 $\mathrm{mW/m^2}$ corresponds to 1.4 $\mathrm{\mu cal/cm^2s}$. This is the heat flow unit ($\mathrm{mW/m^2}$) which may be referred to as HFU. The temperature of a volume of rock is a function of the heat flux and the conductivity of rocks and fluids. The increase in temperature with depth (temperature gradient) is called the geothermal gradient (dT/dz). Typical geothermal gradients may also be written as T and in sedimentary basins are usually 25–45°C/km.

$$\mathrm{Joule} = \frac{\mathrm{kg} \times \mathrm{m}^2}{\mathrm{s}^2}$$
$$1\,\mathrm{Joule} = 0.239\ \mathrm{cal}$$
$$1\,\mathrm{Watt} = 1\,\mathrm{joule/s}$$

In a sedimentary basin there is the background heat flux from the underlying basement, and granitic rocks have higher heat production and temperatures than basic rocks. The sedimentary sequences overlying the basement also produce heat by radioactive reaction and black shales with a high content of organic matter often have a relatively high uranium content. This additional heat source may be significant in terms of increasing the heat flux and the geothermal gradients in sedimentary basins.

The temperature distribution (geothermal gradients) in sedimentary basins can vary regionally and over geologic time. This determines both the generation and expulsion of petroleum, and it also strongly influences the reservoir quality. It is therefore important to understand the processes that control heat transport in sedimentary basins.

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$$\text{The temperature gradient } \nabla T(\mathrm{d}T/\mathrm{d}z) = \text{Heat flow}(Q)/\text{Conductivity}(c). \quad (9.2)$$

In rocks with low conductivity (like mudstones and shales) the geothermal gradients will be high and in highly conductive rocks (like salt) the geothermal gradients will be low (Fig. 9.1). The thermal conductivity of salt (halite and anydrite) is 5.5 $\mathrm{Wm^{-1}{}^\circ C^{-1}}$, while shales may have conductivities between 1.0 and 2.5 $\mathrm{Wm^{-1}{}^\circ C^{-1}}$. Sandstones and limestones have values between shales and salt.

In a situation where a rock is filled with stationary porewater the total heat flux (Q) is the sum of the heat conducted through both the rock's matrix and pores (porosity φ filled with fluids):

$$Q = C_\mathrm{r}(1-\varphi) + \varphi C_\mathrm{f} \quad (9.3)$$

C_r is the conductivity of the solid rock and C_f is the conductivity of the fluids (usually water) in the pore space. Water (fresh) at 20°C has a conductivity of 0.6 W/mK while seawater and saline brines are much more conductive. The conductivity of common sedimentary minerals ranges from 7.7 $\mathrm{Wm^{-1}{}^\circ C^{-1}}$ for quartz to 1.8 for illite and smectite. The conductivity is therefore to a large extent a function of the quartz content and the water content (porosity).

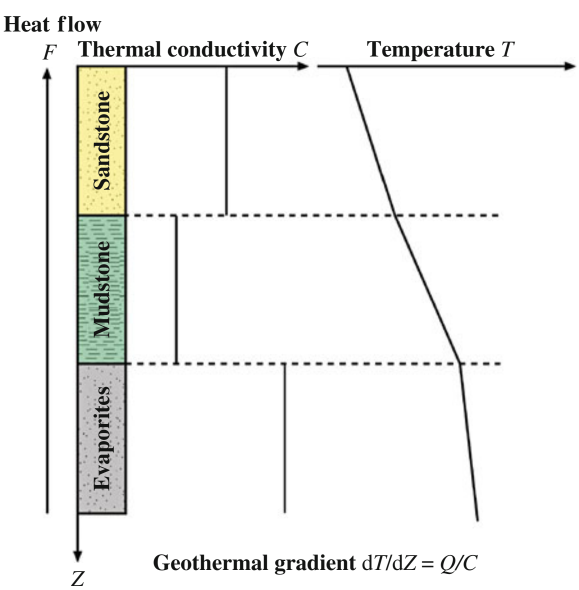

Fig. 9.1 Relationship between heat flow (Q), conductivity (C) and geothermal gradients. The geothermal gradients have an inverse relation with the conductivity for the same heat flux (F). Thick salt layers or domes cause higher geothermal gradients above the salt and low temperatures below the salt

The conductivity of shales from the North Sea ranges from about 0.8 to 1.1 $\mathrm{Wm^{-1}{}^\circ C^{-1}}$ (Midttømme et al. 1997) so they are not very much more conductive than water. The conductivity parallel to bedding may be up to 70% higher than perpendicular to bedding.

Most of the heat transport is vertical except around hydrothermal or igneous intrusions, but in the case of steeply dipping beds the conductivity would be higher.

Over a limited vertical interval of the sedimentary section the heat flow may be relatively constant and we see from Eq. (9.2) that the geothermal gradient is inversely related to the conductivity.

When there are rocks with low conductivity near the surface the geothermal gradient will be higher so that the underlying sediments will be warmer. This is called a blanketing effect. Mudstones with low thermal conductivity on top of granites or older sedimentary rocks will have this effect.

Salt with high conductivity has the opposite effect. Because the temperature gradient through salt is low, the temperature will be relatively high at the top of the salt and low at the bottom. This has consequences for maturation of source rocks. This is a very important effect for petroleum prospects below thick salt layers, i.e. in the Gulf of Mexico, offshore Brazil, West Africa and the North Sea. The temperatures below the salt will be significantly lower than normal at this depth. This means that the reservoir quality of sandstones reservoirs will be better due to less quartz cement. Lower temperatures will also preserve more petroleum as oil or condensate as there will be less cracking to gas.

The heat flux is only constant in an equilibrium situation. When sediments subside they are heated and a part of the background heat flux is used to heat the subsiding rocks (Fig. 9.2). This is equal to the heat capacity of the rocks and the subsidence rate. In basins with high sedimentation rates the heat flow is strongly reduced and in the Plio-Pleistocene depocentres the geothermal gradients are down to 20–25°C/km (Harrison and Summa 1991).

During subsidence and sedimentation the sediments must be heated, and this heat is taken from the background heat flow and the geothermal gradient is reduced.

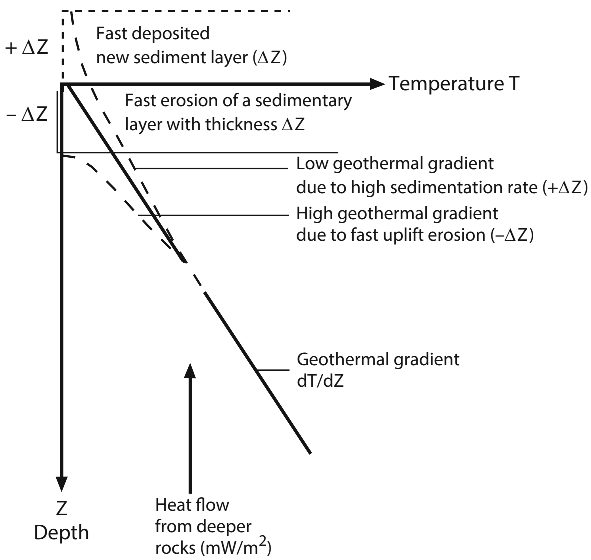

Fig. 9.2 Geothermal gradients as a function of rapid uplift (erosion) or subsidence (sedimentation). During subsidence some of the heat flow is used to heat the subsiding sediments and underlying basement and this will reduce the geothermal gradients, forming cold basins. During uplift the heat from cooling rocks will add to the heat flux, producing steeper geothermal gradients

During subsidence the heat flux is:

$$Q_s = Q_b - F_{s.}\, C_{(hc)} \cdot dT/dZ$$

The heat capacity of rock including porewater is $(C_{(hc)})$.

Here Q_b is the background heat flux from the basement, F_s is the subsidence rate (downward flux of rocks) and C_{hc} is the heat capacity of the rocks. When rocks are uplifted and cooled the heat given off from the cooling rocks adds to the background heat flow:

During uplift the heat flux becomes:

$$Q_s = Q_b + F_{ur} \cdot C_h dT/dZ$$

Here F_{ur} is the rate of uplift. The heat capacity of the mineral matrix may be estimated at about 8–900 J/kgK. The heat capacity of water (C_{hw}) = 4,200 J/kgK.

In terms of volume the heat capacity of rocks is however close to 2,500 J/dm^{3}K.

This means that at 20% porosity about 2.5 times as much heat is stored in the mineral matrix (density 2.7)

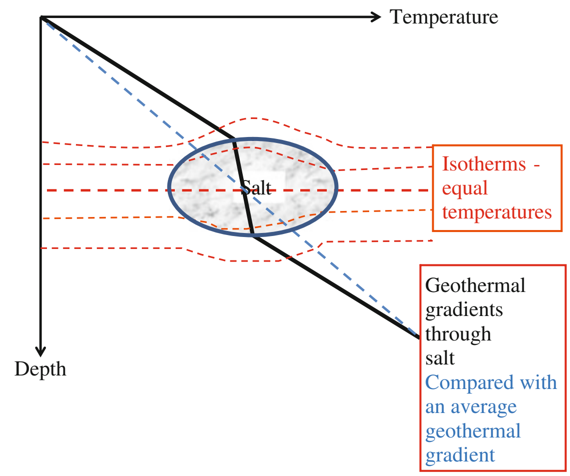

Fig. 9.3 Geothermal gradients are strongly influenced by layers of salt or salt domes. Since the heat flow is relatively constant the geothermal gradient must be low through the highly conductive salt. As a result the overlying sediments will be warmer than normal while the underlying sediments will be cooler

as in the water phase. During the first period of advective flow along a fault or through permeable sandstone beds a high percentage of the advected heat will be lost by conduction to the mineral matrix.

9.1 Heat Transport by Fluid Flow

When fluid, usually water, is transported in a sedimentary basin there is also heat transport unless the transport is parallel to the isotherm (Fig. 9.3).

The advective heat transport Q_t is proportional to the flux of water (Darcy velocity $F = \mathrm{m}^3/\mathrm{m}^2/s$), the heat capacity of water (C_{hw}), and the geothermal gradient (T).

$$Q_t = F_h \cdot C_{hw} \nabla T \cdot \sin \alpha$$

Here α is the angle between the direction of fluid flow and the isotherms which are lines with equal temperature. h is the length along the direction of fluid flow.

During compaction-driven flow the flow rates are in most cases too small for this heat transport to be significant. Focused compaction-driven flow may cause a significant heat flow by advection, but only if the rate of porewater flow is very high. The average water flow upwards relative to the sediments is very low, but relative to the seafloor the porewater is in most cases sinking.

Meteoric water fluxes along aquifers into sedimentary basins are many orders of magnitude faster than in compaction-driven flow; in some cases the downwards flow of cool meteoric water from mountains into sedimentary basins may cause a significant reduction in the geothermal gradients.

9.2 Heat Transported by Conduction and by Fluid Flow (Advection)

The relative contribution from these types of heat transport can be expressed by the Peclet number (Pe):

$$\mathrm{Pe} = \rho_\mathrm{f}\, C_\mathrm{f}\, Q_\mathrm{z}\, L / C_\mathrm{r}(1-\varphi) + \varphi C_\mathrm{f}$$

Here ρ_f is the fluid density, C_f the heat capacity of the fluid, Q_z the vertical component of the Darcy velocity, L the length of the flow path, and C_r and C_f the respective thermal conductivities of the solid phases (minerals) and the fluids (water) (Person and Garven 1992). Sedimentary sequences with permeable sandstones and limestones normally include low permeability shales and siltstones The distance between them tends to control the length of the flow path and the height of the convection cells. In thick permeable sandstones and limestones the vertical flow and transport of heat is faster, resulting in lower geothermal gradients. Bjørlykke et al. 1988. See Chap. 10.

The conductivity of water depends on temperature and salinity but is much lower than that of the mineral matrix (0.6 W/m°C and 2.5–3.5 W/m°C, respectively). Hot porewater therefore rapidly loses its heat to the mineral matrix. Convection is driven by the primary temperature gradients. Porewater convection does change the temperature field but temperature perturbations due to this flow are not very large (Ludvigsen 1992).

Numerical calculations of fluid flow in modern sedimentary basins like the Gulf of Mexico basin show that compaction-driven porewater flow is insignificant in terms of advective transport of heat (Harrison and Summa 1991). In the North Sea basin, too, the geothermal gradients only vary within rather narrow limits (35–40°C/km). The occurrence of locally higher values offshore Western Norway has been attributed to the effect of recent glacial erosion producing transient thermal heat flows (Hermanrud et al. 1991).

In the Mississippi Valley, USA, compaction-driven flow has been shown to be quite insufficient to generate hot fluids capable of precipitating ores (Bethke 1986). Compaction-driven flow from thrust belts may produce significant thermal perturbations on a relatively local scale, but modelling suggests that such flow is insufficient to cause large-scale thermal anomalies in the adjacent foreland (Deming et al. 1990). In continental rifts like the Rhine Graben, where the rift margins are exposed and elevated topographically, groundwater flow can to a large extent explain the observed thermal anomalies (Person and Garven 1992).

9.3 Importance of Heat Flow and Geothermal Gradients

Heat flow is a very important parameter, which strongly influences geothermal gradients and rates of petroleum generation. It also strongly influences rates of quartz cementation and other types of chemical compaction in siliceous sediments.

Heat flow and geothermal gradients are also important for the utilisation of geothermal energy and heat pumps in the ground or in rocks.

In sedimentary basins we have a heat flow from the basement into the overlying sedimentary sequence. The composition of the basement rock determines the rate. Granitic rocks with high potassium and uranium content will produce more heat than rocks like anorthosites and gabbros which are very low in potassium. Offshore Norway the background heat flow varies significantly depending on the basement rocks. Organic-rich shales like the Upper Jurassic source rocks from the North Sea basin may also contribute significant heat because of the radioactivity (high uranium content).

As we have seen above, increasing sedimentation rates will reduce the geothermal gradient because some of the heat flux is used to heat new layers of subsiding sediments. Cold basins with rapid subsidence and low geothermal gradients (<20–25°C/km) require deep burial of the source rocks before they can generate petroleum, both because of low temperature and the short geologic time (<2–3 million years) for petroleum generation. The small time/temperature integral will also result in little quartz cement in reservoir sandstones. The amount of quartz cement can also be used as a measure of the time/temperature index. The temperature history is an important parameter in basin modelling because it influences the sediment density and therefore the rate of subsidence and the generation of hydrocarbons, and also the reservoir quality. The sediment density is increased by porosity reduction during diagenesis and mineral dehydration. Heating however causes a slight expansion of the minerals and density reduction.

The heat flow and the geothermal gradients may change over geologic time and that complicates basin modelling. In subsiding basins, however, it is the geothermal gradients during the last part of the subsidence which are most important. In the North Sea Basin we may have had relatively high geothermal gradients in late Jurassic times, but both the source rocks and the reservoir rocks were then only buried to rather shallow depths and the temperatures were still relatively low, except close to volcanic intrusions and hydrothermal activity.

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Chapter 10

Subsurface Water and Fluid Flow in Sedimentary Basins

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The pore spaces in sedimentary basins are mostly filled with water. Oil and gas are the exceptions and most of the information we have about fluid flow in sedimentary basins is derived from the composition of water and the pressure gradients in the water phase. It is therefore important to characterise and understand the variations in the composition of these waters. All the porewater may be referred to as *subsurface water* but the water that is analysed from exploration wells or is produced during oil production is usually called *formation water* or *oil field brines*.

The composition of subsurface water can provide vital information for several different practical purposes:

(1) The water composition may give information about the origin of the water and the pattern of fluid flow in the basin.
(2) Differences in water composition with respect to salinity or isotopic content may provide important information about communication (permeability and fluid flow) across barriers like shale layers or faults.
(3) The composition of water produced together with oil may give information about the rock units from which the water is being drained. When water injection is applied to a reservoir, the composition of the injected seawater is different from that of the formation water and this can be used to determine if there is a breakthrough of the injected water to the production well.
(4) The water composition determines the density of the water column which is required to interpret and calibrate the fluid pressure data from wells.
(5) Water composition and its resistivity are critical for the calibration of well logs. The resistivity is crucial for calculating the formation factor F, which is the resistivity of the formation water relative to the resistivity of the whole fluid-saturated rock.

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Water buried with the sediments has long been referred to as *connate* water and thought to represent the original seawater. This is an unfortunate term because the origin of such water is very complex and meteoric water has been found to have a much stronger influence in sedimentary basins than earlier assumed. In addition, seawater changes its composition significantly as soon as it is buried; the first major change is the removal of sulphate ions in the sulphate-reducing zone (Figs. 10.1 and 10.2). Then the porewater will gradually approach equilibrium with the minerals present in the sediments as the temperature increases.

Subsurface waters are mainly derived from:

(1) Seawater buried with the sediments.
(2) Meteoric water, which is groundwater (originally rainwater) that can flow from land to far offshore along permeable beds, mostly sandstones and limestones.
(3) Water released by dehydration of minerals i.e. from gypsum, or clay minerals like smectite and kaolinite.
(4) Hydrothermal water introduced by igneous activity. This is also referred to as juvenile water.

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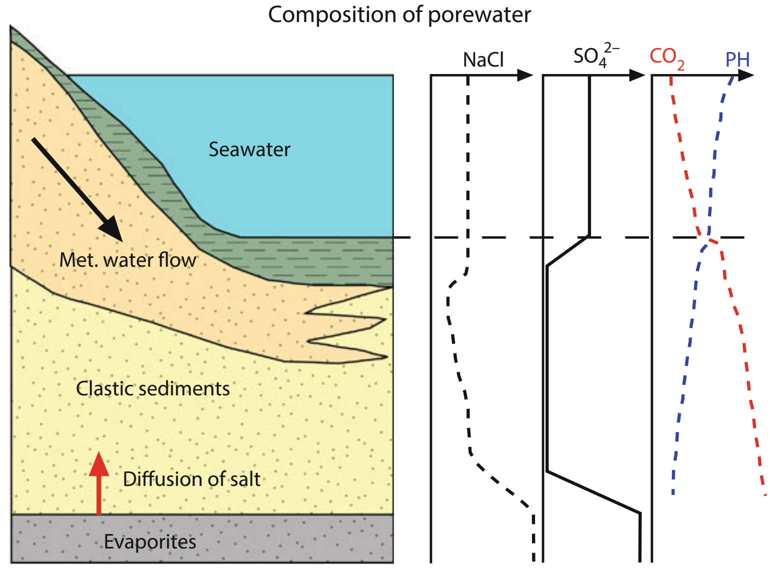
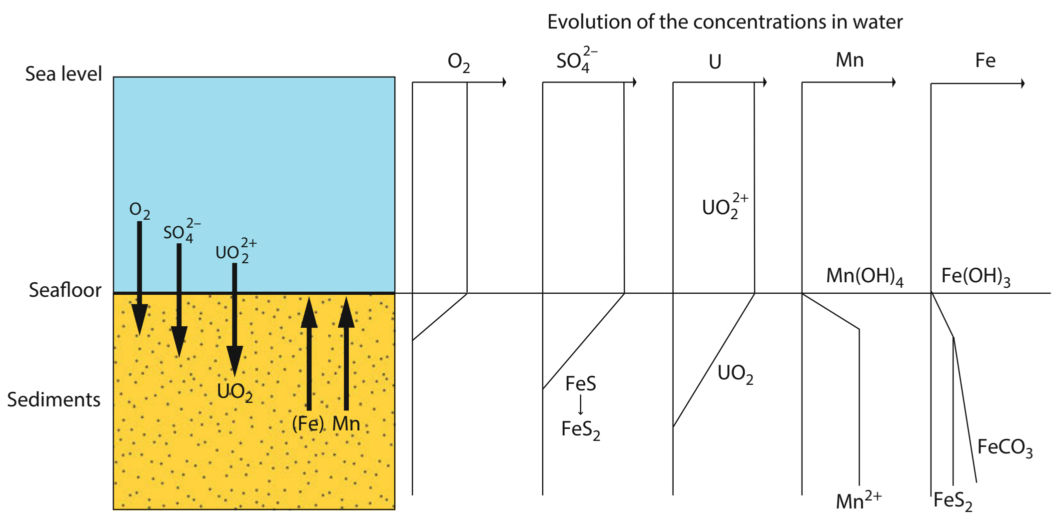


Transport and precipitation of elements near the seafloor occurs mainly by diffusion across the redox boundary.

Fig. 10.1 Simplified illustration of how the composition of porewater is influenced by reaction near the seafloor and supply of freshwater into sedimentary basins from land areas



The redox boundary is a function of the supply of free oxygen or sulphate from the seawater and the consumption of oxygen mainly by oxidation of organic matter.
In the absence of H_2S from sulphur reduction, Fe^{2+} is more soluble but may precipitate as $FeCO_3$.

Fig. 10.2 Only a few centimetres below the seafloor there are important reactions between the seawater, which is normally oxidising, and reducing porewater. Elements that are most soluble in the oxidised state (sulphur and uranium) are concentrated below the redox boundary while elements like iron and manganese are precipitated above the redox boundary on the seafloor

10.1 Composition of Formation Water

In most cases the formation water is saline and Cl^- is by far the dominant anion, so that the Na^+/Cl^- ratio is less than 1. The rest of the positive charge is mostly made up of $Mg^{2+}$and Ca^{2+}. The composition of porewater is a function of its primary origin, modified by the mineral composition of the sediments, the temperature and the quantity of dissolved gases, particularly CO_2.

Pure water of meteoric origin derived from fresh groundwater may gradually mix with more saline water and become more brackish. Meteoric water usually has less than 10,000 ppm dissolved material compared to seawater, which has 35,000 ppm and contains more bicarbonate (HCO_3^-) and small amounts of magnesium, sodium and calcium. Because meteoric water comes from a land surface, it brings with it oxygen and bacteria which can break down hydrocarbons (see "Biodegradation"). In sediments with even small amounts of organic matter the porewater quickly becomes reducing as oxygen is consumed by breakdown of the organic matter. The composition of meteoric water alters as it reacts with the more readily soluble minerals in the sediments. Small amounts of carbonate makes the meteoric water basic and will act as a buffer with respect to pH. Meteoric water will flow from land areas where the groundwater table is elevated above sea level through sandstones serving as aquifers far offshore, depending on the pressure head of the groundwater table.

Drilling on the continental shelves, for example off the east coast of the USA and also offshore Africa, shows that meteoric water aquifers are widespread and that water which is virtually fresh is sometimes found below the seabed as far out as 100 km from the coast. In the northern North Sea there is also isotopic evidence that the formation water in shallow reservoirs is partly of meteoric origin. This also shows that the porewater is often stratified and there is very limited vertical mixing of porewater with different salinities.

Marine porewater, which is present in sediments when they are deposited, will initially have an approximately normal salinity but sulphate is removed by sulphate reduction just below the seafloor. Clay minerals act as ion exchangers, absorbing cations like K^+ and Mg^{2+} from the porewater. Compacted clay and mud may function as a membrane. Clay minerals are normally negatively charged on the surface and particularly at the ends, where there are broken silicate bonds. Negatively charged ions are repulsed and held back by the membrane due to the negative charges of the clay minerals. In order for the charges on both sides to equalise, the small ions, particularly H^+, must move in the opposite direction, and as a result there is a higher H^+ concentration (lower pH). This process, which will concentrate salt, is called *salt sieving*. Membranes also discriminate selectively amongst different cations, depending on their size and charge. Those alkali ions which are strongly hydrated (Na^+, Li^+) and also Mg^{2+}, will to a lesser degree be adsorbed onto the surface of the clays and will be more mobile than for example K^+ and Rb^+, which are less hydrated. At higher temperatures the hydration becomes less effective and ions like Mg^{2+} are more available to be adsorbed on clay minerals and form carbonate minerals like dolomite.

At 70−100°C, smectite will dissolve and form illite and water. At 120–140°C kaolinite will become unstable and form illite, quartz and water. This process binds cations, particularly K^+, and releases pure water, thus reducing the salinity of the porewater. As a result the porewater in shales often has a lower salinity than that in sandstones at the same depth. The composition of porewater in sandstones varies greatly, depending on whether they contain meteoric water or not.

The quantity of dissolved solids in porewater increases as a function of depth in most cases. Concentrations of 100,000–300,000 ppm of dissolved matter (total dissolved solids, TDS) are typical for basins with evaporite beds. With increasing temperature, the kinetic obstacles to mineral solution and precipitation reactions are reduced. At temperatures above about 80°C the porewater will tend to be in equilibrium with most of the minerals present. In sedimentary basins with evaporites, these will greatly influence the composition of the porewater in overlying formations. Dissolved salts move upwards through compaction-driven porewater flow; diffusion due to high concentration gradients also plays a major role. Sedimentary basins along the Atlantic Coast in areas where Mesozoic evaporites are deposited have porewater compositions which are essentially different from those in areas north of this palaeoclimatic belt. The South American continental shelf, the Gulf

Coast and the area off the East Coast of the USA are characterised by Jurassic/Cretaceous evaporites. In the North Sea the extension of Zechstein evaporites forms an important boundary which is reflected in the porewater composition of overlying Mesozoic sediments. In addition to chemical analysis, isotope composition ($^{13}C/^{12}C$ and $^{18}O/^{16}O$) can provide valuable information concerning the origin of the porewater.

Many of the dissolved ions in porewater are in equilibrium with the minerals present. They are then not very useful as indicators of the origin of the water. Cl− and Br− are better tracers for fluid flow as they do not react very much with the minerals present.

The water produced from drill stem tests may be strongly contaminated by the drilling mud filtrate and the composition of the mud should be considered when using the analyses of such waters. Water produced during production is less likely to be strongly contaminated because of the larger volume involved. The results of the analyses may be expressed as % or ppm.

$$\mathrm{mg/l = ppm/density\ of\ water}$$

$$\mathrm{meq/l = mg/l \times valence/mol.weight}$$

The main anions in subsurface waters are Cl−, HCO_3^- and SO_4^{2-}, and the main cations are Na^+, K^+ and Ca^{++}.

Meteoric water is characterised initially by low ionic strength but it then reacts with minerals and also amorphous phases like opal A. Unless meteoric water flows through evaporites the chlorinity will remain very low and the main anions will be bicarbonate (HCO_3) or carbonate $\left(CO_3^{2-}\right)$. Sulphate $\left(SO_4^{2-}\right)$ may form in meteoric water due to oxidation of sulphides in rocks and of sulphur in organic matter, but the sulphate will tend to be reduced to sulphides by sulphate-reducing bacteria.

Porewater of marine origin naturally starts with the composition of seawater but only a few centimetres below the seafloor most of the free oxygen is removed from the porewater due to oxidation of organic matter in the sediments. A few metres below the seabed nearly all the sulphate which was in the seawater has been consumed by sulphate-reducing bacteria. It is therefore a characteristic of so-called connate water that it has very low sulphate content. The chlorinity, however, remains practically unchanged because Cl− is not consumed by any significant diagenetic process.

Steep concentration gradients and a strong drive for transport by diffusion exist across the redox boundary because of the difference in solubility of many ions between the seawater, which is normally oxidised, and the porewater below. Sulphate $\left(SO_4^{2-}\right)$ is transported downwards and is reduced to sulphides below the redox boundary. Reduced sulphur reacts with iron-bearing minerals including iron oxides (haematite) to form pyrite (FeS_2). Reduced manganese and iron (Mn^{2+}, Fe^{2+}) will be transported upwards and precipitated above the redox boundary. While much of the iron will be trapped in the reduced state as sulphides, manganese sulphides are rather soluble and very little Mn will therefore be trapped below the redox boundary. Ferrous iron (Fe^{2+}) may also be trapped as carbonate such as siderite ($FeCO_3$). Manganese is therefore transported upwards more efficiently than iron and may form large deposits (as manganese nodules) in deepwater environments where the sedimentation rate is low.

The composition of subsurface water is strongly influenced by the dissolution of evaporites where these are present but the effect can often be shown to be rather local (radius <1 km). Meteoric water can dilute the chlorinity of porewater but mixing of porewater is not very efficient in sedimentary basins. This is because the flow is slow and laminar and pore waters with different salinities have different densities, which also tends to inhibit mixing. Transport by diffusion from high to lower salinity may nevertheless be significant over distances of a few hundred metres, depending on the diffusion constant of the sediment matrix. Low permeability shales also have low diffusion constants. Dehydration of minerals such as gypsum or clay minerals like smectite, kaolinite and gibbsite also causes reductions in salinity because pure crystal-bound water is released into the porewater.

Higher salinities than seawater are in most cases due to the dissolution of evaporites.

High salinity porewater is found in the central North Sea above the Permian evaporites in the Central Graben, while in the northern North Sea the porewater is of normal salinity or brackish composition. In the northern North Sea where there are no evaporites, formation water salinity is in most cases close to that of seawater, or less saline (brackish) due probably to a component of meteoric water. Near evaporites the porewater is also often rich in calcium due to gypsum or anhydrite, with a corresponding reduction in sodium, so that $CaCl_2$ is an important dissolved salt. Dissolved NaCl may be transported away from the evaporites by diffusion or by porewater flow (advection). In both cases the salinity will be reduced away from the salt deposit and it is unlikely that the dissolved salt would have become sufficiently concentrated for halite to be precipitated again.

The halite and also gypsum commonly observed in small amounts in sandstone cores is due to evaporation in the core store; they are not diagenetic minerals as has sometimes been reported. During tectonic uplift and erosion, hydration of minerals like anhydrite causes increased salinity but this is normally diluted with meteoric water. Near the surface the salinity can of course increase by evaporation. In the seawater at shallow water depth the pH is relatively high because the water is often at least nearly saturated with respect to calcite and the pCO_2 is low. At greater water depth in the ocean and also below the seafloor the solubility of CO_2 increases, lowering the pH. Carbon dioxide is consumed near the ocean surface by photosynthesis and released by oxidation of organic matter sinking towards the ocean floor. Just below the seafloor in the sulphate reduction zone more CO_2 is produced during fomentation of organic matter. Then at greater burial depth, thermal maturation of kerogen releases CO_2. Organic acids are also generated from the source rocks and probably also from the oil. Porewater is, however, like seawater a buffered solution and organic acids are weak acids. Compared to the total buffering system of, firstly, the silicate mineral system and, secondly, the carbonate system, the addition of relatively small amounts of comparatively weak acids (organic acids) will not change the pH of the porewater significantly (Hutcheon 1989). These different types of CO_2 have characteristic ranges in $\delta^{13}C$ composition.

Formation analyses from sedimentary basins show that the pore waters are frequently crudely stratified with respect to salinity and this puts constraints on porewater flow. Also the oxygen isotope compositions may vary with depth, probably at least partly due to diagenetic reactions. At shallow depth and low temperature the porewater is normally out of equilibrium with respect to the silicate minerals because of the slow reaction rates. Meteoric water is usually highly supersaturated with respect to quartz because it does not precipitate at low temperatures ($<$70–80°C). Seawater, on the other hand, is usually very much undersaturated with respect to quartz because silica is taken out of seawater by siliceous organisms, mainly diatoms.

Amorphous silica from organisms (opal A) may survive during burial down to 1.5–2 km (60–80°C) before dissolving. Opal A is replaced first by opal CT, which is unstable and will be replaced by quartz. As long as opal A and opal CT exist in the sediments the porewater is supersaturated with respect to quartz. At higher temperatures the porewater becomes closer to equilibrium with quartz and other minerals present.

The isotopic composition of subsurface water reflects to a large extent the initial composition. Porewater with a marine origin has characteristic values close to Standard Mean Ocean Water (SMOW). More negative values may indicate the introduction of meteoric water into a marine basin. At greater burial depth the composition of porewater is more influenced by the reactions of the minerals. As a general rule dissolution of minerals formed at low temperature, i.e. during weathering (kaolinite, smectite), causes the pore waters to become more positive when they dissolve at greater depth (higher temperatures). This is because the minerals that precipitated at a higher temperature, in this case illite, will contain less ^{18}O. Dissolution and precipitation of clastic quartz, which was originally precipitated at high temperature, will shift the porewater in a negative direction when low temperature quartz, with more ^{18}O, is precipitated.

We now have considerable data on the composition of porewater in sedimentary basins from exploration and production wells. It is clear that the porewater composition varies greatly over distances of a few hundred metres and this is evidence of very limited mixing by advection. Except around salt domes the porewater seems to be crudely stratified with respect to both salinity and isotopic composition $\left(\delta^{18}O\right)$. This puts important constraints on the transport of solids in solution by fluid flow. The saline porewater is not transported very far from the salt. With increasing temperature the density of porewater is reduced, while it is increased by increasing salinity.

10.2 Composition of Porewater in an Oil Field

Analyses of porewater that is present in the oil field along with the oil provide information about fluid flow and migration of oil. Differences in salinity and chemical composition between parts of a reservoir may indicate lack of communication by fluid flow and also diffusion between compartments. The isotopic composition of the formation water may also help to indicate degrees of communication in a reservoir in addition to changes in the composition of oil. Strontium isotopes have been used for this purpose. If there are primary differences in the composition of oil across an oil field this can be used to monitor the contribution from the different parts during production. During onshore production freshwater may be used for water injection into the reservoir. Offshore, seawater is injected, but this has a composition which is distinctly different from the formation water.

At any given time the amounts of solids in solution are very small except in highly saline porewater near evaporites. The solubility of most silicates and also carbonate minerals is sufficiently low that the porewater composition of sedimentary basins is almost totally controlled by the solid phases. Porewater in sedimentary basins often shows evidence of stratification with respect to both salinity and isotopic composition, which precludes large scale mixing of porewater. The composition of the formation water in reservoirs may provide useful information about the source of the water and communication within the reservoir, in the same way as the composition of the oil gives information about the source of the oil.

10.3 Fluid Flow in Sedimentary Basins

Fluid flow in sedimentary basins is important because it determines the distribution of pore pressures in the water phase, and also in oil and gas. High pore pressures may also be a hazard when drilling wells. The fluid phases have the capacity to transport solids in solution, and heat by fluid flow (advection) and by diffusion. The flow of fluids may be through the pore network in the rock matrix or along fractures, and this is a principal difference between these two types of flow. This part of the chapter will discuss the factors controlling the flow of fluids in sedimentary basins.

The fluids are mostly water, but may also be oil and gases including air, filling the pores between the grains in the sediments. The grains are in most cases minerals but some may be amorphous (e.g. silica – opal A, or organic matter and kerogen). *Porosity* is the percentage (or fraction) of the rock volume which is filled with fluids. This may be also expressed by the *void ratio* which is the ratio between the volume of voids (porosity) and solids. What is called *void* is not strictly void but filled with fluids and is the same as porosity. The relation between porosity (φ) and void ratio (V_r) is thus:

$$V_r = 1/(1 - \varphi)$$

In the oil industry porosity is mostly used while rock and soil mechanics literature tends to use void ratio.

Porewater flow in sedimentary basins can be classified according to the origin of the water and the driving mechanism for the flow:

(1) Meteoric water flow is sourced by groundwater (originally rainwater) and in most cases the groundwater table is above sea level, providing a drive downwards into the basin. This water is normally fresh (low salinity) except in very arid regions where meteoric water may dissolve evaporites.
(2) Compaction-driven water is driven by the effective stress and thermally-driven chemical compaction which causes a reduction in available pore space. The upward component of this flow is a function of the rate of porosity reduction (compaction) in the underlying sediments.
(3) Density-driven flow is driven by gradients in the fluid density due to differences in salinity or temperature. Thermal convection is driven by the thermal expansion of water. As the temperature increases downwards in sedimentary basins the density is reduced, creating a density inversion with depth. Flow driven by thermal convection differs from the two other types of flow in that the same water is used over again and is not dependant on an external supply of porewater.

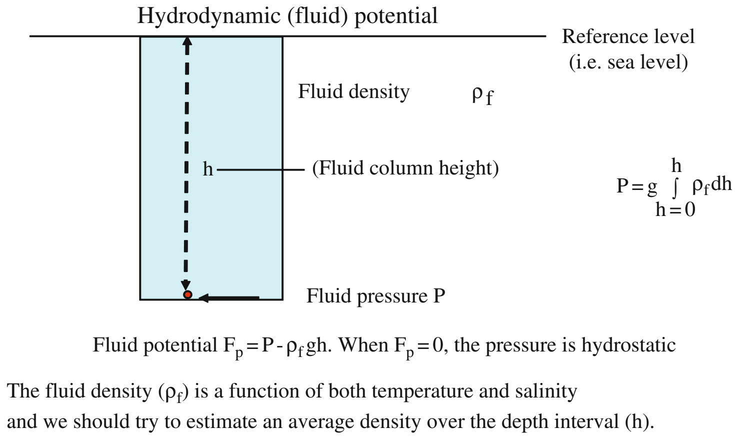


Fig. 10.3 Illustration of fluid potential which is the difference between the fluid pressure at a certain depth and the weight of the overlying column of porewater and seawater

10.4 Fluid Potentials

The flow of fluids in sedimentary basins follows simple fluid dynamics laws. Fluids do not necessarily flow from higher to lower pressures, but from higher to lower fluid potentials. The fluid potential is defined as:

$$\mathrm{F_p} = P - \rho g h.$$

Here P is the fluid pressure, ρ is the density of the fluid, g the acceleration of gravity and h is the distance up to some reference level, which in a sedimentary basin could be the sea level or the water table (Fig. 10.3). The fluid potential is thus the potential for fluid flow, so if the fluid potential is zero there can not be any flow.

The fluid potential is an expression of the *deviation* from the pressure gradient due to the density of the fluid column. In the case of water the hydrodynamic potential expresses the deviation from the hydrostatic pressure gradient ($\rho_w g$), which is defined by the weight of the water column. In the case of groundwater flow the fluid potential is usually referred to as the hydraulic potential, which is the mechanical energy per unit volume of groundwater. The lithostatic stress, which may also be referred to as the total stress, is the weight of the rock column saturated with fluids (ρ_r), for the most part water. The difference between the total stress ($\rho_\mathrm{r} gh$) and the pore pressure (P) is the effective stress (σ_ev) which is transmitted by the sediment particles or the rock:

$$\sigma_\mathrm{ev} = \rho_\mathrm{r} gh - P$$

Differences in hydrodynamic potentials can also be expressed in terms of potentiometric (or piezometric) surfaces which are the heights to which water would rise above sea level (or some other reference datum like the groundwater level) in an open pipe from a rock in the subsurface. Pore waters with a higher potentiometric surface than sea level are defined as *overpressured* while those that have a potentiometric (piezometric) surface close to sea level or the groundwater table are normally pressured.

It is often stated that fluids flow from high to lower pressure but this is obviously not always true. In the ocean the pressure increases from the surface down towards the bottom but water does not flow from the bottom to the surface because there is in most cases no potentiometric head. The pressure gradient in the ocean water is close to the density gradient ($\rho_\mathrm{r} gh$) and ocean currents are driven by very small differences in fluid potential due to changes in temperature.

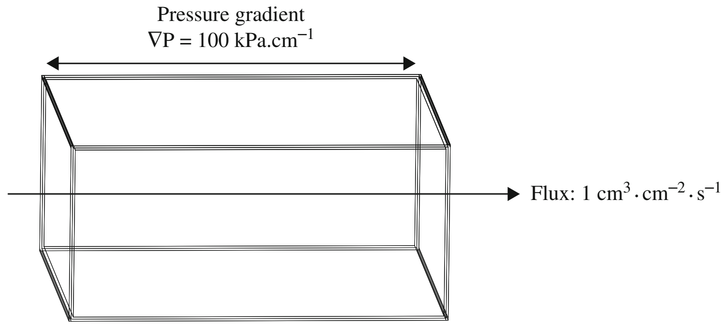
To maintain significant pressure (potentiometric) gradients, there must be a resistance to flow; in the case of flow in porous rocks this is measured as *permeability*. Fluid flow is a function of permeability (k) and viscosity (μ), and the flow of water and other fluids, can be described by the Darcy equation:

$$F = \nabla P \cdot k/\mu.$$

Permeability

The permeability is the resistance to flow in any material.

If the permeability is 1 Darcy (1 D) the fluid flux is 1 $cm^3 \cdot cm^{-2} \cdot s^{-1}$, when the pressure gradient is 1 $atm.cm^{-1}$ (100 $kPa.cm^{-1}$) in the direction of flow and the viscosity of the fluids is 1 centipoise (as for water at 20°C).



SI unit:

1 D = 10^{-12} m^2
1 mD (one millidarcy) = 10^{-3} D (10^{-15} m^2)
1 nD (one nanodarcy) = 10^{-9} D (10^{-21} m^2)

Fig. 10.4 Illustration of the Darcy equation for fluid flow in sedimentary basins

The flux F can be expressed as volumes of fluids passing through a certain cross-section in a given time, i.e. $cm^{-2} \cdot cm^{-2}s^{-2}$ (Fig. 10.4). This is also called the *Darcy velocity*. The *absolute velocity* of the flow is $v(cm/s) = flux(cm^3 \cdot cm^{-2}s^{-})$ divided by the porosity (φ). So if the porosity is 0.1 (10%) the velocity is 10 times the flux. In reality the velocity is a little higher because the fluid pathway is not along a straight line. Since the porosity varies greatly along the flow path it is more useful to use the flux (cm^3/cm^2) or the Darcy velocity (cm/s) as a measure of fluid flow rates. ∇P is the potentiometric gradient, i.e. the change in potential over a certain distance. In the horizontal direction this is the same as the pressure gradient.

Given a pressure P_1 in a point X_1 at a depth h_1, and a pressure P_2 in a point X_2 at the depth h_2, the *potentiometric gradient* (∇P) is:

$$\nabla P = ((P_1 - \rho g h_1) - (\mathrm{P}_2 - pgh_2))/X_2 - X.$$
$$(X_2 - X_1 \text{ is the distance between } P_1 \text{and } P_{2.})$$

The *permeability k* is an expression of the resistance to flow and is a constant in the Darcy equation which relates solely to the properties of the rock. Permeability has the dimension of $m^2(1\ \text{Darcy} = 10^{-12}\ m^2)$. The permeability of a rock may be referred to as the *absolute or intrinsic permeability* to make it clear that it only relates to the characteristics of the rock, as opposed to the *relative permeability* which is the permeability of one immiscible fluid in the presence of another fluid, compared to the permeability with 100% saturation of one fluid.

The *hydraulic conductivity* (K) or *transmissibility* is an expression of the ability of the rocks to conduct or transmit fluids of a certain viscosity (μ). The conductivity has the dimension of m/s or ft/s and can be calculated from the permeability if the viscosity is known ($K = k \cdot g/\mu$). At about 20°C, the kinematic viscosity of water is $1 : 10^{-6}\ m^2/s$ and at 100°C it is $0.210^{-6}\ m^2/s$ which is one centipoise. 1 Darcy (permeability) $(k) = 10^{-5}\ K$ (conductivity) for water (strictly $9.66 \times 10^{-6}\ K$), 1 Darcy = $10^{-12}\ m^2$ or 10^{-8} cm^2. Well sorted and poorly cemented sand may have permeabilities between 1 and 10 Darcy. In tight shales the permeability is typically below 1 nanodarcy (10^{-9} Darcy) or even much lower (Fig. 10.4).

Water is not very compressible. The compressibility is close to $4.4 \times 10^{-10}/Pa^{-1}$. The pressure of a 1 km water column (10 MPa) causes a compression of water of about 0.4% or 4 m. The expansion of water during tectonic uplift or release of overpressure is therefore relatively small. Usually the cooling of

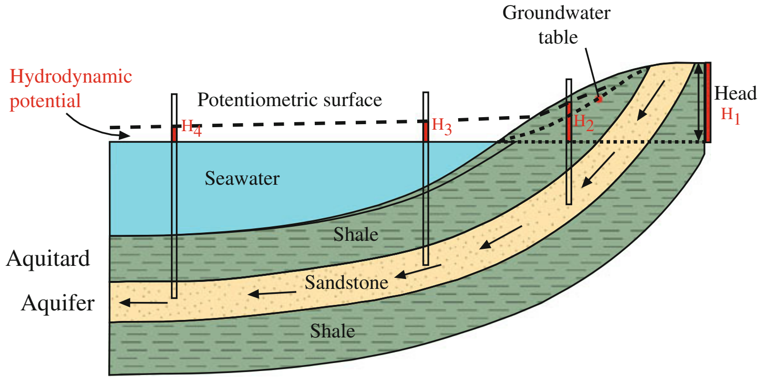


Fig. 10.5 Fluid flow along a sandstone which is a confined aquifer overlain by a shale which is an aquitard

water associated with uplift will cause a thermal contraction which is greater than the expansion due to pressure reduction, so that the porewater becomes denser.

Porewater flow is oriented perpendicular to points of equal hydrodynamic potential (isopotentional lines) because that will represent the steepest potentiometric gradient. That will be the case in relatively homogeneous sediments but generally the flow is very much controlled by the distribution of highly permeable *aquifers* such as poorly cemented sandstone layers, and of shales which serve as low permeability *aquitards* (fluid barriers) (Fig. 10.5).

The distribution of very high permeability aquifer sandstones and low permeability aquicludes (mostly shales, salts and cemented layers in salts) that control most of the flow in sedimentary basins is closely linked to primary sedimentary facies, tectonic developments and diagenesis. These depositional and diagenetic processes determine the "plumbing system" in the basin and this must be the starting point for fluid flow modelling.

Aquicludes are beds of such low permeability that they can not transmit significant quantities of fluids under normal hydraulic gradients. These are groundwater hydraulics terms but over geological time even low permeability shale will transmit some fluid. Some shales may have extremely low permeability and be practically impermeable, thus capable of maintaining overpressures on this timescale. The term *seal* is often used as equivalent to aquiclude in the oil industry, and may describe a seal for both oil and water. Shales which are not completely sealing with respect to water may nevertheless prevent oil from entering

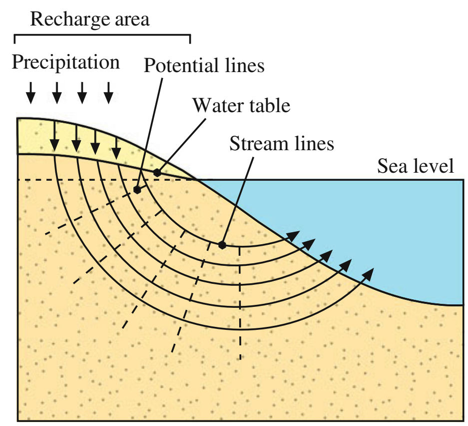


Fig. 10.6 Flow of meteoric water in massive sand, perpendicular to the pressure distribution (potential lines)

into the small pores, due to capillary forces. Fractures may be open and provide highly permeable pathways, but others may be almost impermeable barriers to fluid flow because they are closed or cemented up. In a massive sand the flow is very different and controlled by the orientation of the isopotential lines which may be controlled by the groundwater table (Fig. 10.6).

Fluid flow modelling that uncritically applies the Darcy equations to fluid flow in sedimentary basins can lead to quite unrealistic results. Frequently the porewater flux is calculated from an observed pressure gradient and from assumed or modelled permeabilities for the different lithologies. One of the main problems with such calculations is that the permeability can not be determined very accurately. It typically varies by several orders of magnitude from bed to bed up through a sequence and there may also be large variations along the bedding due to sedimentary structures.

In the case of flow *perpendicular* to the bedding the average permeability is the *harmonic mean* of the permeability of the individual beds. The harmonic mean of the permeabilities of a number of beds (n) is defined by:

$$k_{\mathrm{h}} = \left[i/n \sum_{i=1}^{n} 1/k_i \right]^{-1}$$

The harmonic mean is strongly influenced by the bed with the lowest permeabilities such as a tight shale or a carbonate-cemented interval. For flow *parallel* to bedding the arithmetic mean is relevant. Even if we measure the permeability at relatively short intervals in a cored section it is very difficult to come up with a good average permeability, partly because relatively thin, low permeability, layers affect the value to such a degree.

In exploration we want to make predictions ahead of drilling, but without core data it is very difficult to provide assumptions about the permeability distribution with the degree of confidence needed for modelling fluid flow. In most cases, however, the fluid flux (F) is primarily constrained by the supply of fluids. The flux of meteoric water (groundwater) flow is limited by the infiltration of rainwater into the ground. The potentiometric gradient near the surface is the slope of the groundwater table (or the potentiometric surface). If the rainfall is 1 m/year the infiltration may be 0.3 m/year and this is the initial flux in the recharge area. If this flux is continuous for 1 million years the total flow is then $3 \times 10^5\ \mathrm{m}^3 \cdot \mathrm{m}^{-2}$ which is very significant, several orders of magnitude larger than the average compaction-driven flow. However the meteoric water flux is greatest near the surface and decreases rapidly with depth. We must remember that the meteoric water flowing into the basin also must flow up to the surface. Sandstones that pinch out in mudstones or shales will support only a very low flux despite the high permeability of the sandstones.

Fluid flow resulting from differences in hydrodynamic potential can be treated mathematically using Darcy's equation and mass conserving equations during flow (continuity equations). Modelling flow for whole basins is very complex. The orientation and distribution of permeable sediments (usually sandstones and limestones) will to a large extent dominate the pattern of fluid flow. At depths greater than 3–4 km (100°C) quartz cementation may be more extensive along the fault planes than in the adjacent sandstones.

The importance of fault planes as conduits for fluid flow is greatest in well-cemented uplifted sedimentary rocks or in metamorphic rocks, because the matrix permeability is so low. These are rocks that have been subject to unloading and usually some extension (fracturing), and they possess high rock strengths which can prevent fractures from closing. Deformation of well-cemented rocks may produce rock fragments (brecciation) which by wedging the faults may help to resist horizontal stress, keeping the fractures open. In more porous sedimentary rocks like sandstones and limestones, which are not well cemented and have higher matrix permeabilities, fractures are less critical. In subsiding sedimentary basins the sediments do not usually have sufficient strength to resist the horizontal stress that is trying to close any open faults and fractures. Faults in this setting are therefore more likely to be barriers than conduits for fluid flow (see Chap. 11 on rock mechanics).

Prediction of fluid pressure ahead of drilling in the basin depends on the permeability distribution in three dimensions over distances of several kilometres. Even if the geology is known in great detail, it would still be difficult to specify sufficient details about the permeability to obtain a realistic fluid flow model based on sedimentology and structural geology. During exploration we normally have insufficient data to model fluid pressure. During production much more data is available on the distribution of permeabilities and the model can be constrained by the pressure response to production, and to water injection.

10.5 Meteoric Water Flow

If the permeability is homogeneous in all directions, the porewater flow can be calculated from the elevation of the groundwater table and the fluid densities alone. The flow is perpendicular to lines with equal potentials. In an isotropic rock matrix with constant fluid density, the flow of meteoric porewater follows a curved pattern perpendicular to the isopotential lines (Fig. 10.6). Sedimentary rocks are generally very inhomogeneous and the contrasts in permeability caused by clay layers and sand or gravel beds will generally totally dominate the flow of water (Fig. 10.5). While sand and gravel beds may have permeabilities of 1–10 Darcy, the permeability of poorly-compacted mud may be 0.01 mD or lower. In such cases, mathematical modelling is of little value if the stratigraphy, sedimentology and structural deformation of the sedimentary sequences are not interpreted correctly. Modelling can nevertheless help to constrain some of the interpretations by calculating the consequences of different alternatives.

Meteoric water (freshwater) has a potentiometric head defined by the groundwater table, which is normally higher than sea level. Fresh porewater will thus flow into marine sedimentary basins from the coastlines (Fig. 10.7). Along the coast and underneath islands, lenses of freshwater float on more saline porewater like an iceberg in the sea. The depth to which freshwater will penetrate is a function of the density difference between the freshwater and the saline water:

$$D = H\rho_{\mathrm{mw}}/(\rho_{\mathrm{sw}} - \rho_{\mathrm{mw}})$$

Here, D is the depth of penetration below sea level, H is the height of the groundwater table above sea level and ρ_{sw} and ρ_{mw} are the densities of saline water and meteoric pore waters, respectively. For $\rho_{\mathrm{sw}} = 1.025\,\mathrm{g/cm^3}$ and $\rho_{\mathrm{mw}} = 1.0\,\mathrm{g/cm^3}$, $D = 40\,H$, meaning that the depth of the freshwater wedge is 40 times the height of the groundwater table underneath an island or within a confined aquifer along the coast. If the meteoric water becomes brackish by mixing with saline waters, the density difference is reduced and the depth of penetration can be deeper because $(\rho_{\mathrm{sw}} - \rho_{\mathrm{mw}})$ becomes smaller. The depth of meteoric water flow into highly saline porewater is very much less. Relative sea level changes serve as a pumping mechanism, driving meteoric water into sedimentary basins at low sea level stands due to the increase in hydrodynamic head. In this way early diagenesis may be linked to sequence stratigraphy.

During the last glaciations sea level was lowered more than 100 m several times. This increased the head of the groundwater table on the land area, pushing meteoric water deep into sedimentary basins. Under completely hydrostatic conditions, a 100 m head (10 MPa pressure) should theoretically correspond to a penetration down to about 4 km following the above calculations. However, the flow will mostly follow permeable beds (aquifers) and may extend a great distance out from the coastline. Well log analyses from offshore Georgia, USA, suggest that freshwater extended for up to 100 km offshore just a few metres below the seafloor (Manheim and Paull 1981). In the modern Gulf of Mexico Basin, the depth of meteoric water penetration is estimated to

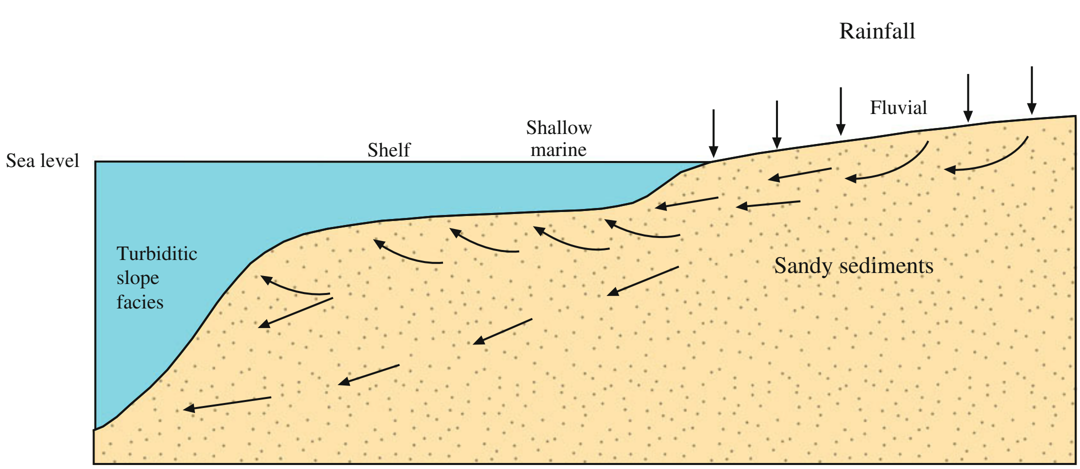


Fig. 10.7 Flow of meteoric water into sedimentary basins. The fluid flux will decrease away from the coastline and with increasing depth.

be about 2 km (Harrison and Summa 1991). The depth of penetration does not depend only on the head of the meteoric water but also on compaction processes in the sediments, which can generate overpressures that may exceed the meteoric water head. Even slight overpressures due to compaction will strongly reduce the depth of meteoric water penetration.

The rainfall, the catchment area and the percentage of infiltration into the groundwater determine the upper limit of meteoric water flow. Meteoric water which flows down into a sedimentary basin must eventually flow up to the surface, in order to maintain a continuous flow. Permeable sandstones are hydraulically almost dead ends if they pinch out into very low permeability mudstones. Pressure will then build up in the aquifer and reduce the meteoric water inflow, forcing water to flow through the overlying mudstones. We must remember that even if the fluid flux is small, the large area of contact between a sand layer which serves as aquifer and the overlying mud will allow relatively high volumes of porewater to escape upwards through the mud. Even if the flow per area (flux) is small through the mudstones the area for vertical flow is large compared to the vertical cross-section. At shallow depths (<500 m) prior to severe compaction, overpressure is rarely developed as the permeability is much higher than in compacted mudstones and shales.

The degree of meteoric water flushing is highly dependent on climate and facies. The land surface and vegetation determine the percentage of rainfall which infiltrates down to the groundwater. Fluvial, deltaic and nearshore shallow marine sediments will be flushed by meteoric water shortly after deposition. The flux is then likely to be high and the porewater is still very much undersaturated with respect to feldspar and mica. The total volume of water flowing through each volume of sediment is inversely related to sedimentation rates. At high sedimentation rates the sediments spend less time in the zone of meteoric water flushing. Sands deposited in more distal shelf facies (nearer the shelf edge) and turbidites (on the slopes) are normally less well connected to the main groundwater wedge, so that the flux is lower and the porewater is closer to equilibrium with respect to the mineral phases.

In the North Sea basin it has been demonstrated that reservoir sandstones deposited in fluvial and shallow marine environments have been subjected to more feldspar dissolution (secondary porosity) and contain more authigenic kaolinite than sandstones representing turbidite facies (Bjørlykke and Aagaard 1992). Sediments in sedimentary basins like the North Sea may be intensively flushed by meteoric water immediately after deposition and also after uplift episodes and erosion. When tectonic uplift results in subaerial exposure and the formation of islands, meteoric water is collected on land and driven into the subsurface around the islands and adjacent to other land areas. The sediments most strongly affected by meteoric water leaching are constantly being removed by erosion, however. This may explain why there is not always much evidence of feldspar leaching and high kaolinite contents immediately below unconformities.

Good examples of clay mineral diagenesis related to modern groundwater systems have been observed down to 3–400 m depth in the Mississippi Gulf coastal plain (Hanor and Mcmanus 1988). In Canada there is isotopic evidence of recent meteoric water diagenesis extending several hundred metres below the land surface (Longstaffe 1984). It must be stressed that the isotopic composition of the porewater acquires a meteoric signature as soon as a volume of meteoric water has displaced the marine (connate) porewater. New minerals (like calcite, kaolinite and quartz) precipitated in this porewater will reflect that isotopic signature.

10.6 Porewater Flow Driven by Thermal Convection

Thermal convection is an effective mechanism for the mass transfer of dissolved material in sedimentary basins, because the same water can be used over and over again (Wood and Hewett 1982, Davis et al. 1985). The limitation of fluid (water) supply, which constrains compaction-driven flow, is then eliminated. Thermal convection may occur because the density of water decreases with depth in a sedimentary basin as the temperature increases, due to the thermal expansion of water. This creates an inverse density gradient which may be unstable. If the isotherms (lines of equal temperature) are horizontal, the density as a function of temperature will not vary horizontally and there is no flow unless the water overturns. The denser upper layers of porewater may start to overturn and

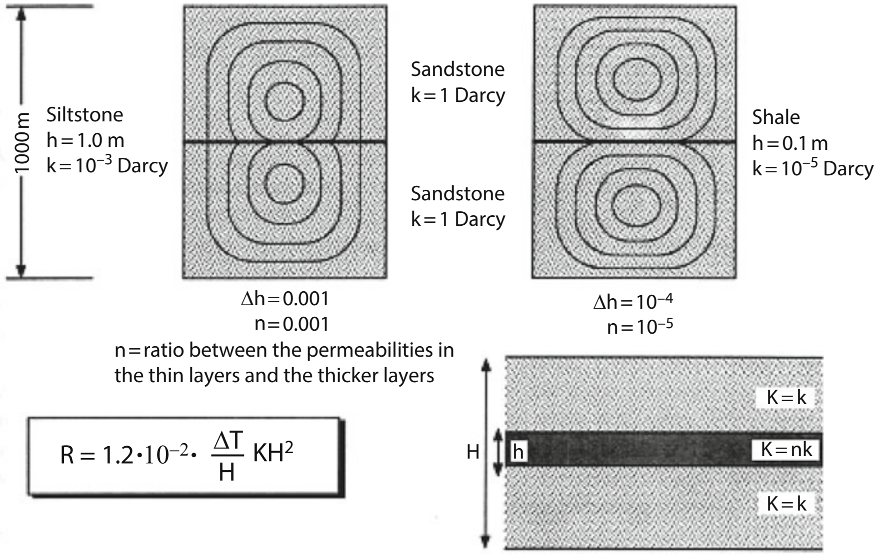


Fig. 10.8 A modelling of the low permeability layers on vertical Rayleigh convection in sedimentary basins (Bjørlykke et al. 1988)

sink into the less dense water below. The condition required for such overturning can be expressed in terms of a critical Rayleigh number. In the case of thermal convection of water the Rayleigh number can be defined as follows (Bjørlykke et al. 1988):

$$R = g\beta\Delta THk/\kappa\mu$$

Here, g is acceleration due to gravity, β is the coefficient of thermal expansion of the fluid (water), ΔT is the temperature difference between the upper and lower boundaries of the convection cell, H is the thickness of the layers, k is the permeability, κ is the thermal diffusivity and μ is the viscosity. Assuming reasonable values for the properties of water the equation can be expressed in a simpler form (Bjørlykke et al. 1988):

$$R = 1.2 \times 10^{-2} k\nabla TH$$

The critical Rayleigh number which must be exceeded for Rayleigh convection to occur is about 40. We see that the most critical factors are the height of the water column and the permeability of the rocks. If the permeability is 1 Darcy and the geothermal gradient 30°C/km, the thickness (H) of the permeable layer (sandstone) must exceed about 300 m for the critical Rayleigh number to be exceeded so that thermal convection can occur. Sedimentary rocks, though, are rarely uniform and the vertical permeability typically changes abruptly in a sequence of sandstones and shales. Thin layers (0.1 m) of low permeability shales or cemented layers in sandstones may cause almost complete flow separation, producing smaller convection cells instead of potentially larger ones (Fig. 10.8) (Bjørlykke et al. 1988). Each of the convection cells may then have insufficient height (small H) to exceed the critical Rayleigh number. The low vertical permeability in layered sequences suggests that Rayleigh convection is probably not very important in sedimentary basins. Several hundred metre thick sandstones with no thin shales or cemented intervals are rarely encountered.

Non-Rayleigh convection will always take place when the isotherms are not horizontal. This is because the temperature and the fluid density are then not constant in the horizontal direction. This situation is always unstable because there is a potentiometric drive for the waters to overturn, which will produce some fluid flow without the need to exceed a critical Rayleigh number. The velocity for non-Rayleigh convection is:

$$v = g \cdot k \cdot \beta \sin \alpha \nabla T/\mu$$

When the geothermal gradients vary only moderately within a basin, the slope of the isotherms (α) will be small and the flow velocity very low. Sloping isotherms will also result from sloping beds because the heat flux is reflected when the conductivity of the beds varies, but this effect is also normally quite small. From the equation above we see that the flow rates are a function of the isotherm slope, as well as a function of the height of the convection cell which often is equal to the thickness of a sandstone bed. The thickness of the sandstone beds (H) or the distance between the low permeability shales, will normally define the height of the convection cells and in most sedimentary sequences the distance between thin shales or even clay laminae is only a few metres or less. Although non-Rayleigh convection nearly always occurs to some degree it is probably rather insignificant in most cases in terms of the transport of solids in solution in sedimentary basins. This is because the low slopes of the isotherms and the low vertical permeability result in very low flow rates, mostly inside rather thin sandstones separated by shales. These are probably not very significant in terms of solid transport in connection with diagenetic processes. Around igneous or hydrothermal intrusions, however, the lateral change in geothermal gradients and the slope of the isotherms may be very high and then thermal convection is very important. Also around salt domes geothermal gradients may change due to the higher conductivity. Inverse salinity gradients will increase the drive for thermal convection, while normal salinity gradients will make the porewater more stable. Even moderate salinity gradients strongly influence fluid flow in sedimentary basins (Fig. 10.9).

The increase in density due to the salinity may totally or partly offset the density reduction due to the thermal expansion of water. At a salinity gradient of 30 ppm/m, and an average geothermal gradient, the effects of the thermal expansion of water are more than offset, so that the water becomes denser with depth (Fig. 10.9). This effectively removes any drive for convective flow (Bjørlykke et al. 1988). When such trends are recorded in formation water analyses or well logs it provides strong evidence that vertical mixing is not taking place, because convection would have destroyed the salinity gradients (Gran et al. 1992). Around salt domes the permeabilities are often rather low, further reducing the potential for fluid flow.

The salinity in sediments surrounding salt diapirs can be used to trace fluid flow, since Cl^- is not consumed to any significant degree by diagenetic reactions. Analyses of the salinity distribution around salt domes from offshore Louisiana show some evidence of convection, but the observed salinity stratification and the lack of more mixing and dilution of the saline porewater suggest this convection is very slow (Ranganathan and Hanor 1988, Evans and Nunn 1989). Inverted salinity gradients are only likely to develop around salt diapers or underneath salt layers. The inverse salinity gradients can only sustain flow on the down-going limb of a convection cell. Unlike thermal convection there is no mechanism to make the water flow up again and the water would also gradually homogenise with respect to salinity. In basins like the North Sea, porewater analyses

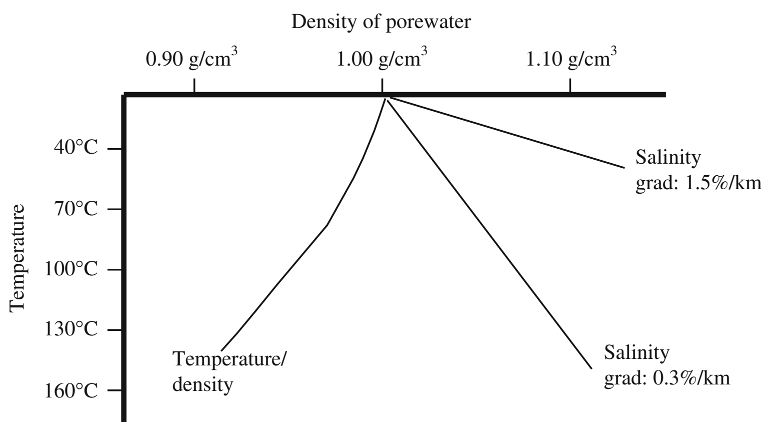


Fig. 10.9 Density of water as a function of salinity and temperature (from Bjørlykke et al. 1988). The increase in temperature causes a thermal expansion and a density inversion while salinity gradients may have the opposite effect

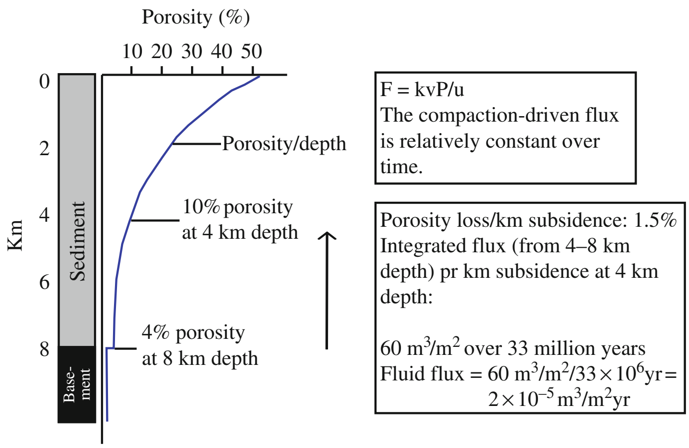


Fig. 10.10 Illustration of a porosity/depth function in a sedimentary basin. The integrated area defined by this *curve* is an expression of the total volume of water in the basin and the *slope* (the derivative) is an expression of the compaction-driven water flux from each layer

demonstrate that the porewater is at least stratified in a crude way with respect to its composition. The salinity increases downwards towards the Permian salts, effectively ruling out large scale convection and excursions of compaction-driven flow from the deeper parts of the basin into the overlying sequence (Gran et al. 1992).

There is also a trend towards more positive $\delta^{18}O$ values with depth, which again confirms some degree of porewater stratification (Moss et al. 2003). The vertical salinity gradients and the isotopic composition of the porewater confirm that the porewater is not undergoing convection on a large scale and that there is no large flow of porewater from the deeper part of the basin to shallower depths. This has important consequences for diagenetic models in connection with fluid transport of solids in solution.

10.7 Compaction-Driven Porewater Flow

As sediments compact they lose porosity and the excess porewater has to be expelled. This is the driving force for compaction-driven flow. The rate of porosity loss is a function of effective stress, lithology, temperature and time. The porosity/depth functions observed in sedimentary basins may be very complex, depending on the lithologies. At the transition between two lithologies the porosity may increase with depth but for a uniform lithology the porosity will decrease with depth. If we integrate the porosity/depth function from the seafloor though the sedimentary sequences to the underlying basement we obtain an area A below the porosity/depth curve. This is an expression of the total volume of water in the basin per unit area (Fig. 10.10). The total compaction-driven flow of water in the basin is a function of the changes in the porosity/depth curve. As new layers of sediment are deposited the underlying sediments compact and the porewater is forced upwards.

It is possible to show that at a constant sedimentation rate the average upward component of the compaction-driven flow is always equal to or lower than the subsidence rate (Caritat 1989). The porewater is therefore moving upwards through the sedimentary sequence but nearly always downwards relative to the seafloor. We may say that the sediments are sinking through a column of porewater. Typical sedimentation rates in sedimentary basins are 0.1–0.01 mm/year and the average rates of upwards porewater flow are lower than these values. Assuming there is no flow of water from the basement there is practically no upward flow in the basal layer of sediments. The porewater in this layer subsides at almost the same rate as the basement. Higher up in the sequence the compaction-driven flow receives contributions from more and more layers. In the uppermost layer the porewater flux is equal to the total integrated porosity loss over time in the underlying sequence. The upwards-flowing porewater is filling the pore space of new layers deposited on the seafloor, and during continued subsidence there is normally no porewater flow up though the seafloor into the water column except when there is local focusing of the flow. During periods with no sedimentation (hiatus), the deposited sediments continue to compact and porewater flows across the redox boundary just below the seabed and dissolved ions like Fe^{2+} and Mn^{2+} may precipitate as $Fe(OH)_3$ and $Mn(OH)_4$.

Very high degrees of focusing are required to obtain high flow rates. It is then easy to understand why compaction-driven flow is several orders of magnitude lower than meteoric water flow, at least in the shallow parts of the basin near land. The slow porewater flow rates also imply that this water has time to approach thermal and chemical equilibrium with the minerals. Porewater will always transport some heat but, compared to the background heat flow by conduction, this is very small and in most cases can be ignored (Bethke 1985, Ludvigsen 1992).

Modelling compaction-driven flow in the Gulf Basin, Harrison and Summa (1991) found that the maximum rate of the vertical component is 2 mm/year (2 km/million years), which is approximately equivalent to the maximum sedimentation rate. This flow rate is too slow to contribute significantly to the heat flow. The temperature distribution in the Gulf Basin at the present day does not reflect compaction-driven porewater flow. Most of the heat transport is by conduction. The average compaction-driven flow within a basin is more or less independent of the permeability because it is a function of the rate of loss of porosity in the underlying sediments. During mechanical compaction, low permeability sediments may result in overpressure and a certain reduction of fluid flow, but at greater depth where the compaction is mostly chemical, the fluid flux is independent of the permeability although variations in permeability may focus the flow.

10.8 Constraints on Water Flow in Sedimentary Basins by Porewater Chemistry

Many sedimentary basins contain evaporites, often occupying the basal part, having formed during the initial rifting. Very high salinity is then typically found in a zone of a few hundred metres adjacent to the salt. This is the case with the Zechstein salt in the North Sea basin. The shallower parts of a basin and in particular near tectonically uplifted areas, may contain porewater of meteoric origin with very low salinity. The isotopic composition of the porewater often shows a very clear stratification. Oxygen isotopes are often negative in the shallow parts of the basin due to the inflow of meteoric water, while they may be positive at greater depth due to diagenetic reactions.

During subsidence there are generally no open fractures because of the ductile properties of subsiding sediments (Bjørlykke and Høeg 1997). The permeabilities in shales are very low, probably less than a nanodarcy (Leonard 1993). Samples measured in the laboratory may show erroneously high values because of fracturing resulting from unloading during core retrieval. Increasing the effective stress during laboratory measurements to compensate for this has been shown to lower the permeabilities by two orders of magnitude (Katsube et al. 1991). In sediments which have been subject to uplift, however, fracture permeability may be important. Laboratory measurements of shale permeabilities may be several orders of magnitude larger than the results of well tests in the field (Oelkers 1996).

In subsiding basins the effective permeabilities on a large scale must be low, not above about 10^{-9} Darcy, to maintain overpressures over time. Calculating the porewater flux due to compaction and using observed pressure gradients, the effective permeability of thick shale sequences can be calculated. Using data from Haltenbanken offshore Norway this method gave permeabilities between 10^{-9} and 10^{-10} Darcy (Olstad et al. 1997). These calculations are based on one dimensional models and the results are most probably minimum values for the permeability values of the cap rock since lateral drainage is not included. Lateral flow would reduce the vertical flow and the permeability would have had to be lower to maintain the observed vertical pressure gradients. Increasing the vertical fluid flux by focussing the flow would imply that the permeability was higher.

10.9 The Importance of Faults

Faults may greatly affect fluid flow in sedimentary basins and may serve either as conduits or barriers, depending on the situation. A clear distinction must be made between flow along, and across, the fault plane. Faults are also important because they may offset

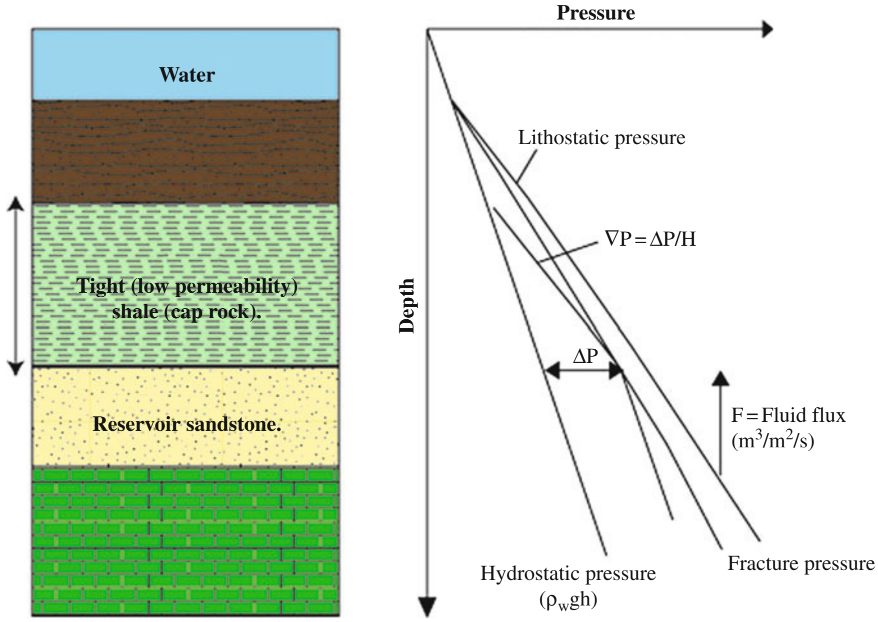


Fig. 10.11 Illustration of fluid pressure and rock pressure (lithostatic). If the permeability in shales (seals) is low enough overpressure will build up. The pressure can not exceed the fracture pressure which is equal to the horizontal stress, as the seal will then leak

porous sandstones (aquifers) against shales, rendering permeable sandstones a dead-end in terms of fluid flow. Faults that cut through sandstones may have a clay smear from adjacent shales or from mica or authigenic kaolinite inside the sandstone, and this may significantly reduce the permeability, in some cases sufficiently to form an oil trap.

Flow along fault planes requires that they are kept open to some degree. A force equivalent to the horizontal stress acts on the fault plane, trying to close it; an equivalent overpressure is required to counteract this in order for the fracture to remain open (Fig. 10.11). However, this corresponds closely to the fracture pressure and even without the presence of a fault the rocks would fracture. At such high overpressure there are very low effective stresses and the sediments are unable to compact mechanically. We must also consider the source of the fluids. When there is no compaction (porosity reduction) in the adjacent sediments, water can not flow from the rock matrix into the fractures. Well-cemented sedimentary rocks and basement rocks have high shear strength and may produce rock fragments (brecciation) during faulting. These rock fragments may wedge the fault plane, helping to resist the horizontal stress. Brecciated fault planes may therefore be important conduits for fluid flow but over time the permeability will gradually be reduced by cementation.

The cements will in most cases not be due to precipitation from advective flow but form by diffusion from the adjacent rock matrix. Carbonate and silicate minerals next to the faults are under lithostatic stress and therefore more readily soluble than when unstressed. In the case of carbonate cement this is fairly clear because upwards (cooling) flow will dissolve calcite rather than precipitate it, due to its retrograde solubility. Brecciated faults contain broken rock fragments and quartz grains that are good nucleation sites for quartz cement formation. A thermodynamic drive towards dissolution of the minerals under stress is therefore likely, with ensuing precipitation in the fault plane. Renewed fracturing is therefore required for the faults to remain permeable.

10.10 Seismic Pumping

In crystalline and well-cemented sedimentary rocks tectonic shear may result in an extensional shear failure which opens up fractures. Increases in the tectonic stress may reduce the opening of such fractures and produce fluid flow that is often referred to as seismic pumping. This mechanism may work in metamorphic and well-cemented sedimentary rocks where the rock strength is high enough to keep fractures from being closed by tectonic stress. In these types of brittle rocks most of the water is present in the fractures and little in the rock matrix. Softer sediments are by contrast more ductile in their response to tectonic stress and fractures will normally not stay sufficiently open to transmit fluid rapidly. In compacting (normally consolidated) sediments most of the water is in the sediment matrix and even if fractures should remain open the rate-limiting step is the flow of porewater from the matrix, which often consists of low permeability mudstones and shales.

If the low permeability seal overlying or surrounding an overpressured part of the basin is broken, i.e. through faulting, a rapid pulse of porewater flow upwards may follow. However due to the low compressibility of water ($4.3 \times 10^{-10}\mathrm{Pa}^{-1}$) the upward flow of porewater necessary to reduce the pressure is relatively small. For an overpressure of 10^7 Pa (potentiometric surface 1 km above sea level), the average expansion of the water would be $4.3{\times}10^{-3}$ (Leonard 1993). In a vertical column through an overpressured sequence this would result in an average upwards flow of 4.3 m for each km of sequence, if all the overpressure was released at one time. If the flow was focused through a smaller cross-section, this figure would increase proportionally.

It is very unlikely that large volumes of shales would reduce their overpressure over a very short time, given their low permeabilities, so the potential for episodes of rapid flow of compaction water is limited. If the water is saturated with respect to gases like carbon dioxide or methane the compressibility of the pore fluid will increase significantly. A reduction in the pressure then will cause gas to come out of solution and form a separate phase, which has a high compressibility. The degree of overpressure is reduced by porewater flow through a leaking seal, but only by a small amount. The pressure will not drop much below fracture pressure before the fine fractures close. In the relatively shallow section this small increase in effective stress may result in some mechanical compaction. Chemical compaction ($>100°$C), however, occurs at a very slow rate which is mostly a function of temperature and time and is relatively independent of the pressure changes related to fracturing. Compaction will therefore slowly build up the pore pressure again unless there is continued flow from the overpressured section.

As we saw from the calculations, the flow resulting directly from the pressure release is rather limited. The main expulsion of porewater is due to sediment compaction, which is an indirect consequence of the reduction of overpressure and increase in effective stress (effective stress = overburden stress minus pore pressure). Compaction and expulsion of porewater resulting from increased effective stress is a gradual and rather slow process and this strongly influences the rate of water supply to the faults from mudstones. The permeability of the surrounding mudstones further limits the rate of flow into the fault zone. If a fault plane does not extend up to the surface (or seafloor), the upwards-moving porewater will have to be accommodated in shallower strata. Large volumes of porewater can not suddenly be injected into shallower sandstones even if the latter are normally pressured.

Flow into shallow aquifers of limited extent will result in a temporary pressure build-up before the water can be displaced, and this will reduce the flow rate. The highest flow rates can be expected when the fault extends all the way to the surface so that the porewater can escape into the water column or onto the land surface. In many sedimentary basins like the North Sea basin, most of the faults extend only into the Cretaceous or lower Tertiary section. These faults were therefore not very active during the early Tertiary, and certainly not during the later Tertiary and Quaternary. The timing of faulting must be taken into account when faults are called upon to explain fluid flow and diagenetic reactions in sedimentary basins. In a sequence with a high clay/sand ratio, faults may be sealing between two sandstones due to the clay smear on the fault plane.

10.11 Episodic Flow

When the source of the fluids is hydrothermal the flow may be episodic, at least when considering individual fractures or a limited area, because new fractures develop and close. Igneous intrusions (sills and dykes) may cause boiling and episodic flow. For a cooling batholith as a whole the fluid flow may be more uniform, depending on the rate of cooling and on the circulation of groundwater by thermal convection. It has also been argued that compaction-driven flow may be episodic when the fracture pressure is reached. Fluids flow through the fractures produced by hydrofracturing but as they escape, pressure will be reduced and the fractures will be reduced or close. It is not so clear if the fracture will stay closed for some time and then leak again or whether it will continue to transmit fluids at a variable rate corresponding to the rate of compaction. At depths greater than about 3 km where most of the compaction is chemical, the rate of compaction will be slow and relatively independent of changes in the stress field. When considering larger volumes of rocks they can not be heated or cooled rapidly, at least not in a non-hydrothermal environment, because of the high specific heat capacity of the rocks. The rate of compaction and the resulting fluid flow will then be relatively uniform over a limited time.

10.12 Formation of Overpressure (Abnormal Pressure)

Overpressure is a term used for subsurface pressures that significantly exceed the hydrostatic pressure. This implies that the flow of porewater to the surface during compaction is resisted to a considerable degree, so that the pressure gradients are increased in the least permeable part of the sediments. Overpressure may be produced by different mechanisms. The simplest type of overpressure is due to the pressure (head) of an elevated groundwater table connected to the basin through an aquifer. Rainwater infiltration into the ground will then help to maintain the pressure even if the aquitards (seals) do not have very low permeability. The overpressure will nevertheless decrease away from the area of recharge. This is also expressed by decreasing piezometric surfaces along the direction of flow. For overpressure to develop due to compaction the permeability in the seal must be many orders of magnitude lower because the fluid flux is very much lower than in the case of meteoric water flow (Bjørlykke 1993). In a meteoric water aquifer the flux may be up to 0.1–1.0 m/year, while the compaction-driven flux is usually less than the sedimentarion rate which may typically be 0.1 mm/year.

During compaction of sediments there must always be a slight overpressure because there must be sufficient pressure gradients for the excess porewater to flow out so that the porosity can be reduced. The Darcy equation shows that the pressure gradient must be an inverse function of the effective permeability of the rocks forming the seal. When we have very low permeabilities a high pressure gradient will build up, which will drive the water out. The term *disequilibrium compaction* has been used to describe the development of overpressure because the permeability is too low for the water to be expelled at lower pressure gradients. A disequilibrieum is always required for compaction to take place, but if the permeablities are not very low, only a slight overpressure is required for the explusion of porewater during compaction. The build-up of overpressure reduces the effective stress with the effect of stopping or at least reducing mechanical compaction (Fig. 10.11). Overpressure thus provides a negative feedback on mechanical compaction. Chemical compaction in siliceous rocks involving quartz cementation will still continue as a function of temperature also during uplift as long as the temperature exceeds 70-80°C but at a lower rate. The pore pressure can then build up to fracture pressure.

From the Darcy equation we see that the pressure gradient (P) is:

$$\nabla P = F \cdot \mu / k$$

It is clear that increases in the fluid flux (F) could cause high overpressure, but the pressure gradient is very sensitive to variation in permeability.

High sedimentation rates and basin subsidence will increase the compaction-driven fluid flux (F) and will contribute to the build-up of overpressure if the permeability is considered to be constant. As shown above, the average upwards flow of porewater is equal to the integrated change in the porosity/depth curve in the underlying sediments.

In addition there is a fluid flux driven by the release of crystal-bound water which in sediments with high contents of water-bearing minerals may be significant. When porewater is heated the thermal expansion of water will also add to the fluid flux but calculations show that this is not very significant.

This is partly because the porewater is not heated very much during basin subsidence since it is moving upwards as the basin and the sediments are sinking.

When solids (e.g. kerogen) are transformed into fluids like oil and gas, a volume expansion may occur. Cracking of oil and the formation of gas also involves a phase change, which will cause increased pressure because of the expansion of gas.

Generation of oil from kerogen gives an increase in volume but calculations suggest that it is not very large because this leaves some solid material (coke) remaining. Generation of gas will cause a higher volume increase, especially at relatively shallow depths.

Even if the total increase in volume is moderate the convertion of solid kerogen to fluid petroleum is a very efficient mechanism for increasing the pore pressure. This is because the fluid/solid ratio is changed and this may cause source rocks to hydrofracture so that the petroleum is expelled. An illustration of such a phase change can be observed in the spring when lenses of ice frozen in the ground melt. The conversion from solid ice to water represents a reduction in overall volume but may generate overpressure because the fluid/solid ratio has been very much increased.

When considering larger compartments in sedimentary basins, however, the pressure is mainly controlled by the water phase which is much more abundant.

Development of overpressure depends on the fluid flux in relation to the permeability of the rocks. The permeability of shales which may serve as seals for overpressure compartments varies greatly and is difficult to predict. A change from 1.0 nD (10^{-9}D) to 0.1 nD will increase the pressure gradient 10 times. We must also remember that for vertical flow perpendicular to bedding the effective permeability is the harmonic average of the permeabilities in the different layers. Thin layers with very low permeability may therefore control the flow rate and the build-up of overpressure. It is therefore very difficult to model and predict overpressure. High overpressures will reduce the effective stress and make the sediments less consolidated. There will then be very little mechanical compaction and compaction-driven fluid flow to build up an overpressure. Fluid transfer from greater depth may cause higher overpressure at shallower levels, though. The onset of chemical compaction, and in particular quartz cementation at temperatures higher than 80–100°C, will reduce the porosity and permeability not only in the sandstones but also in the shales. High sedimentation and subsidence rates will increase the compaction-driven flux but the rate of permeability reduction in the shales (seals) due to chemical compaction will be slower so the permeability of the sealing shales will be higher. This could compensate for the higher fluid flux. Many shales are almost impermeable.

Extremely low permeabilites are required to maintain overpressures in basins that are uplifted and no longer undergo compaction. Overpressured reservoirs in onshore basins in North America have, for the most part, not subsided since the early Tertiary and it is remarkable that the overpressures have been retained without more recent compaction. In the Anadarko Basin the Missisippian and Pennsylvanian sequence is overpressured close to fracture pressure, while the underlying Ordovician rocks are normally pressured. In the Powder River Basin, Cretaceous shales are highly overpressured but not to fracture pressure. Widely distributed free gas (Surdam et al. 1994) may also reduce the permeability for water in the shales. It is not clear if these pressures are maintained by active gas generation at depth. Reservoirs flanking mountain chains like the Rocky Mountains may also be overpressured, here due to meteoric water flow from the mountains.

In sedimentary basins the permeabilities are very much higher parallel to bedding than perpendicular to bedding, and overpressure usually depends more on the lateral drainage than on variations in the vertical permeabilities. Modelling 1 D vertical flow is therefore not very realistic. Faulting that offsets permeable sandstones against tight shales may contribute to the development of overpressure. Synsedimentary growth faulting in particular is very common in basins with high sedimentation rates like the Gulf Coast basins. This is one of the reasons for the widespread overpressuring in such basins.

Models for the prediction of overpressure are often based on changes in fluid flux with less emphasis on the permeability which is more difficult to constrain. This is because the vertical flux depends on the harmonic average of the permeabilities in all the layers also within shales. If the permeability is kept constant, it is possible to model overpressure as a function of other variables such as rates of compaction, hydrocarbon generation and thermal expansion of the porewater. These variables must, however, be compared with the range of permeability values that are likely to exist in a sedimentary basin. Modelling of fluid pressures and the build-up of pore pressure are often based on permeability distributions derived from porosity distributions, which are also poorly constrained. Smectitic mudstones have very low permeabilities even at shallow depth and are particularly effective seals and may be a significant factor causing overpressure. In the North Sea smectite-rich Eocene and Oligocene mudstones and associated sandstones are frequently overpressured at just 1–2 km depth.

The permeability (k) of fine-grained sediments may be related to porosity (φ) and will therefore decrease during compaction. This relationship may be expressed as $k = c\varphi^5$ (Rieke and Chillingarian 1974). It is clear that rather small variations in porosity can produce large variations in permeability. The most important factor in the sediment composition is the surface area (s), which is closely linked to grain size (d). This is expressed in the Konzeny-Carman equation:

$$k = c\varphi^3/(1-\varphi)^2 s^2$$

The effect of the surface area can also be expressed in terms of tortuosity (t):

$$k = \varphi^3 d^2/72\ t(1-\varphi).$$

In the case of smectitic clays the specific surface may be several hundred m^2/g while kaolinite and illite typically have about 10 m^2/g (Skjeveland and Kleppe 1992). The specific surface of mudstones rich in smectite may be more that 10 times that of mudstones with mostly kaolinite, chlorite and illite. According to the Konzeny-Carman equation, the permeability in smectite-rich layers may thus be lower by a factor of 10^{-2} compared to other mudstones. The validity of this equation is however not clear for such fine grained sediments.

10.13 Summary

The origin of porewater in sedimentary basins may be seawater, meteoric water (freshwater) or water released from minerals by dehydration. Pore waters change their composition by reacting with minerals and amorphous phases and approach equilibrium with the mineral phases present at a rate which is kinetically controlled. Highly soluble ions like chlorides, however, are not in equilibrium with the minerals except within evaporite deposits with halite (NaCl). Fluid flow in the deeper parts of sedimentary basins is constrained both by the pressure gradients and the supply of fluids by compaction (reduction in porosity) and by mineral dehydration. Meteoric water is the most important supply of fluids and because it is renewed by rainfall the flow can be maintained for a very long time at shallow depth in sedimentary basins. If other factors are constant the total meteoric water flow through sediments are the inverse of the subsidence rate.

Compaction-driven flow is limited by the volume of water buried in the basin and fluids produced *in situ* by mineral dehydration and petroleum generation. The upwards flow of porewater is usually lower than the sedimentation rate so that the porewater is moving downward relative to sea level. High flow rates can therefore not be sustained except by extreme focusing of the flow.

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Chapter 11

Introduction to Geomechanics: Stress and Strain in Sedimentary Basins

Knut Bjørlykke, Kaare Høeg and Nazmul Haque Mondol

At shallow depths in sedimentary basins there are soft clays and loose silts and sands, unless there are carbonates or carbonate cemented layers. At greater depths they are transformed by diagenetic processes to hard claystones, shales, siltstones and sandstones. Sedimentary rocks continuously undergo physical and chemical changes as a function of burial depth, temperature and time, and important hydro-mechanical parameters change during burial, erosion and uplift. An understanding of these processes is important in order to predict the magnitude and distribution of sediment properties and stresses in the basin. The *in situ* stress condition affects the rock response (strain) to changes in the stress field due to drilling and petroleum production.

Soil and rock mechanics (geomechanics) have mainly been developed to solve engineering problems in relation to landslides and surface and underground construction. These are usually at very shallow depths compared to that of a petroleum reservoir. We will here focus on some aspects of geomechanics of particular relevance for the petroleum geologist.

Porosity loss (volumetric compaction) with time due to increased effective stress is referred to in the engineering literature as *consolidation*, while compaction at constant effective stress is usually called *secondary compression* or *creep*. Compaction in deep sedimentary basins has occurred over geologic time scales at very low strain rates, and at higher temperatures than shallow sediment compaction. Therefore, in addition to mechanical compaction, there are important effects of mineral grain dissolution, precipitation and cementation. This process is called chemical compaction. Here the rate of compaction is controlled by the rate of chemical reactions involving dissolution and precipitation of minerals. Compaction determines the porosity, density and permeability of the sediments which are essential input for basin modelling; petroleum reservoir quality is dependent on the porosity and permeability. The processes of mechanical and chemical sediment compaction (diagenetic processes) determine the physical properties and are also important for understanding seismic velocity records and seismic attributes in sedimentary basins.

11.1 Subsurface Fluid Pressure and Effective Stress Condition

A distinction should be made between total stress, effective stress and fluid pore pressure. This is not always done in technical reports and publications related to petroleum geology.

11.1.1 Total and Effective Stress

In general, stress (σ) is defined as force per unit area. The overburden weight of the sediment including the weight of the fluid in the pore space produces a

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vertical stress (σ_v). For a sedimentary basin with a fairly horizontal surface, and without major lateral variations in the sediment compressibility, the vertical stress at any point can simply be computed as:

$$\sigma_v = \rho_b g h \qquad (11.1)$$

where ρ_b is the average sediment bulk density (solids + fluids) of the overlying sequence, h is the sediment thickness and g is the acceleration of gravity. This is the *vertical total stress* or the *lithostatic stress*. It may be calculated more accurately by integrating the varying density over the depth of the sediment column. The sediment density varies with the density of the grains, mainly minerals, the porosity and the density of the pore fluids which could be water, oil or gas. The effective vertical stress (σ'_v) is defined as the difference between the vertical total stress (σ_v) and the pore pressure (u):

$$\sigma'_v = \sigma_v - u \qquad (11.2)$$

This is the *effective stress* which is sometimes called the average intergranular stresses because it is transmitted through the grain framework. It is the effective stress that governs the mechanical compaction of sediments where little chemical compaction (cementation) has taken place. Stress (σ) is force (F) divided by the area of contact (A). In coarse-grained sediments like sandstones the area of contact may be very small and the stresses between the grains rather high. It should be noted that the local intergranular particle-to-particle contact stress is many times higher than the effective stress as defined here, due to the small area of contact. The total overburden weight is carried by the mineral grain framework and the pore pressure (Fig. 11.1).

The effective stress in the horizontal direction is defined as total horizontal stress minus the pore pressure. The horizontal stress will in general not be equal to the vertical stress as discussed in Sect. 11.3 below. However, the pressure in the pore fluid (pore pressure) is the same in all directions.

The bulk density (ρ_b) of sedimentary rocks varies as a function of the porosity (φ), the density of the fluid (ρ_f) in the pore space, and the density of the solid phase (ρ_m) which is comprised mainly of minerals:

$$\rho_b = \varphi\rho_f + (1 - \varphi)\rho_m \qquad (11.3)$$

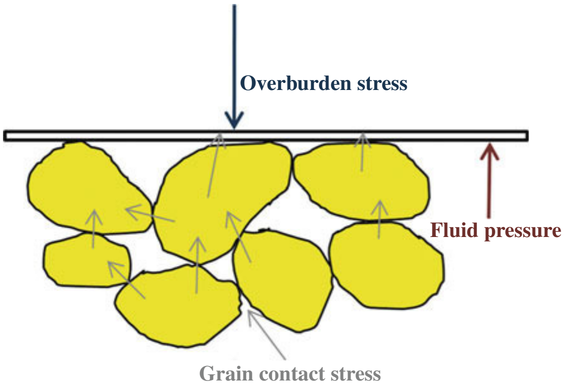

Fig. 11.1 The total vertical stress from the overburden (σ'_v) is carried by the mineral grain framework (solid phase) and the pore pressure (fluid phase). The effective stress is defined as the overburden vertical stress minus the pore pressure and it is transmitted through the grain contacts

The solid phase may also have variable density due to different mineral composition, and in some cases amorphous phases also play a role. Usually the density of the mineral matrix in sandstones and shales is close to 2.65–2.70 g/cm^3. If there are significant contents of denser minerals such as siderite or pyrite the bulk density will be higher. Smectite and mixed-layer minerals have variable but generally lower densities. The fluid density also varies with the composition of water and petroleum. In the case of gas-saturated rocks the bulk density becomes significantly lower. The increase in total vertical stress per metre of depth is commonly called the *lithostatic stress gradient* (Fig. 11.2). At about 10% porosity (and assuming pores filled with water) the lithostatic stress gradient is typically 25 kPa/m (25 MPa/km) corresponding to a mineral density of about 2.66 g/cm^3 as in quartz and illite. At 30% porosity the bulk density of sediments is typically 2.1 g/cm^3.

The rock density is critical for modelling isostasy and backstripping and it is mostly a function of the degree of compaction since mineral densities normally do not vary greatly even if the mineral composition does. Carbonates, particularly dolomite, are however significantly denser than shales and sandstones.

11.1.2 Fluid Pressure

In general, the pressure in the pore fluid at any given point may be equal to the weight of the water column to sea level or groundwater table. The porewater may

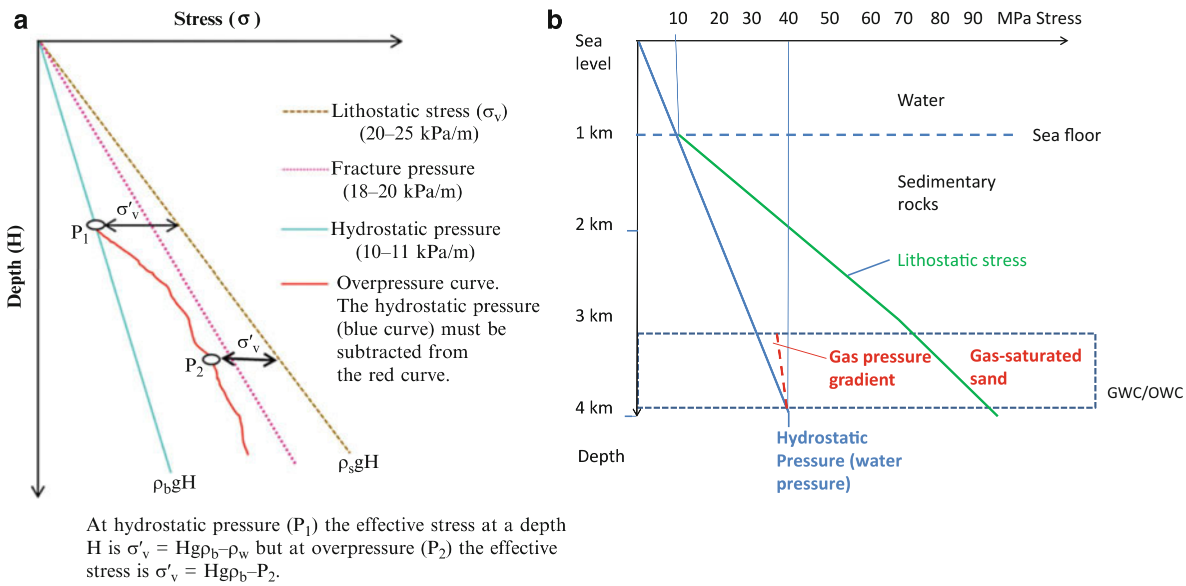

Fig. 11.2 (**a**) Simplified diagram showing the increase in vertical total stress (lithostatic) and hydrostatic pressure as a function of depth. In reality these lines are not strictly straight because the total vertical stress varies as a function of the sediment bulk density (ρ_b) which tends to increase with depth. The hydrostatic pressure curve is a function of the density of the formation water (ρ_w) which varies with temperature and salinity. (**b**) Diagram showing the distribution of stress and fluid pressure in a basin with 1 km water depth. The lithostatic stress is equal to the weight (density) of the overlying sediments and is not a straight line. The hydrostatic pressure is the weight of the water column and the porewater under normal pressure conditions. In a layer saturated with oil or gas the pore pressure is reduced because of the buoyancy relative to water. At the gas/water contact or oil/water contact the pressure in these fluid phases is the same

then be said to be in a hydrostatic state. When significantly higher pressure is measured it is called overpressure, and underpressure also exists in some basins. There may also be a pressure gradient in the porewater which is different from the hydrostatic, and then there will be a fluid flow which is a function of the permeability and the viscosity. The flow may be relatively constant over long time, but if the pore pressure changes over a relatively short time the flow is said to be transient. This would be the case with fluid flow related to earthquakes.

In offshore sedimentary basins the pore fluid pressure (u) measured in oil or water is a result of several contributions:

$$u = \rho_p g H_p + \rho_w g H_w + \rho_{sw} g H_d + \Delta u \qquad (11.4)$$

where $\rho_p g H_p$ is the pressure contribution from a petroleum column of height H_p with the density of petroleum ρ_p, $\rho_w g H_w$ is the pressure due to a water-saturated sequence (H_w), and $\rho_{sw} g H_d$ is the pressure contribution of the seawater column (H_{sw}) density ρ_{sw}. Δu is the *overpressure* (or sometimes *underpressure*) which is any deviation from the *hydrostatic pressure* (Fig. 11.2). Overpressure is sometimes also called abnormal pressure. If u is equal to the hydrostatic pore pressure, then Δu is zero, and the sediment sequence is said to be normally pressured. There is then no tendency to fluid flow.

The overpressure (Δu) can be expressed as the equilibrium height (ΔH) of a hypothetical water column above the level corresponding to hydrostatic pressure. This is the level that water would rise to in a pipe from the formation to the surface. It is called the piezometric or potentiometric surface level. As discussed in Chap. 10 on fluid flow, differences in fluid potentials or piezometric surfaces are expressions of the driving forces for fluid flow in sedimentary basins.

The fluid densities referred to in the equations above are the average densities for a certain fluid column. The density of water decreases with depth due to the temperature increase, but near evaporites the salinity gradient can offset the thermal expansion so that the water becomes denser with depth. The fluid pressure can be calculated more accurately by integrating the fluid density over the height of the fluid column. If the water density is constant and equal to 1.0 g/cm^3, the hydrostatic pressure gradient is 10 kPa/m (0.1 bar/m). Expressed in *psi* (pounds per square inch) the equivalent gradient for freshwater is 0.434 psi/ft. In basins like the North Sea and the Gulf Coast the water density varies significantly, and typical Gulf Coast pressure gradients are 0.465 psi/ft or 10.71 kPa/m (Dickey 1979).

If a standpipe (well) is installed in a normally pressured sediment (no overpressure) the water would rise in the pipe to the sea level or groundwater table. This is called a piezometric or potensiometric surface. Artesian overpressures may be due to meteoric water flow from, for instance, a mountain lake into a sedimentary basin. If fluid pressure is higher than the fluid pressure corresponding to the weight of the fluid above, the water in the standpipe would rise ΔH m above the local water table depending on the degree of overpressure.

During burial and basin subsidence (compaction, compression) the pore pressure is above hydrostatic (transient overpressure) and the water flows out of the sediments as they compact. Unless the permeability in the sediments is very low, only very small overpressures are required for the expulsion of water during compaction. Then the rate of porewater flow is a direct function of the rate of compaction (porosity reduction). If there are low-permeability barriers to flow in all directions, high overpressures may develop during burial because it takes a long time for the pore pressures caused by the added overburden to dissipate/drain. There are also other processes that may lead to overpressure. High sedimentation rates will cause higher rates of compaction and compaction-driven flux. Overpressure will retard mechanical compaction because the effective stress is reduced. The porosity reduction is however very much a function of time and temperature in the case of chemical compaction. Since chemical compaction in siliceous sediments is mainly a function of temperature, compaction will continue even at high overpressures and reduced effective stress.

Lower than hydrostatic pressures (underpressure) can also develop but are less common and are usually formed during uplift in the sedimentary basin. Below are listed three ways in which underpressure may develop:

(1) Tectonic extension may slightly increase porosity and create fractures which need to be filled with fluids, thus lowering the fluid pressure. As water with no gas bubbles has low bulk compressibility (4.10^{-4} MPa^{-1}), a small increase in the porosity caused by the creation of new fractures will produce a significant lowering of pressure. During uplift the sediments no longer compact and water can therefore not flow in from the rock matrix to fill the fractures without lowering the pressure. Extension during uplift will thus tend to draw in meteoric water from above, but if the fractures are not connected so that the water can flow up to the surface, the flow will be rather limited. Compressional tectonics or strike slip tectonics may produce episodes of rapid fluid flow along fractures, often referred to as *seismic pumping*.

(2) Condensation of gas to liquid petroleum may cause reduced fluid volume and lower pore pressure. This may, however, often be compensated for by the expansion of dry gas and the release of gas from porewater.

(3) Cooling and contraction of water (the opposite of aquathermal pressuring) may cause lowering of fluid pressure below hydrostatic.

11.2 Normally Consolidated Versus Overconsolidated Sediments

A layer in a sediment sequence that never before in its geological history has been subjected to higher vertical effective stress than at present, is called normally consolidated (NC). If, on the other hand, the sediment has been subjected to higher effective stresses, e.g. by previous glacial loading, by higher overburden that subsequently has been eroded, and/or by pore pressures in the past that were lower than at present, the sediment is called overconsolidated (OC) as it has been preloaded. The ratio between the past maximum effective vertical stress and the present stress is commonly called the overconsolidation ratio (OCR).

At relatively shallow depths in a sedimentary basin (less than 2–3 km, <70–90°C), the mechanical compaction processes dominate over the chemical compaction in siliceous sediments. At higher temperature (deeper burial) chemical compaction processes become dominant in controlling the rate of compaction. Carbonate sediments may, however, become cemented and highly consolidated at shallow depth. Since this consolidation (compaction) is not due to mechanical compaction but to chemical processes it is sometimes referred to as “pseudo overconsolidation”. Since we refer to the maximum effective stress sediments can not be undercompacted. Poorly compacted sediments have either been subjected to low effective stress or they have a low compressibility.

The hydro-mechanical properties at shallow depths may be very different for a normally consolidated sediment sequence compared with an over-consolidated one, depending on the magnitude of the OCR. For the overconsolidated sediment, the compressibility and permeability are usually much lower and the shear strength significantly higher. As discussed below, the lateral stresses in overconsolidated sediments may be higher than in normally consolidated sediments.

11.3 Horizontal Stresses in Sedimentary Basins

Knowledge of the magnitude and distribution of horizontal stresses in sedimentary basins is important in relation to petroleum exploration, drilling and production. Their magnitude is also important in the interpretation of seismic signals used in field exploration and in reservoir production management. In a sedimentary basin the geomechanical properties vary from those of loose cohesionless sediments at shallow depths to dense and cemented sedimentary rocks at greater depth. This affects the horizontal (lateral) stress distribution with depth.

While the vertical stresses are determined by vertical equilibrium (Eq. 11.1), the magnitude of lateral stresses cannot be determined by equilibrium equations and is statically indeterminate. Their magnitudes are governed by a number of factors, including the overburden/erosion (loading/unloading) and uplift history of the basin and the deformation characteristics of the sedimentary rocks. These are a result of gravitational and tectonic forces, and also of stress changes caused by chemical compaction and the accompanying volume change. Their magnitude is determined based on an understanding of the geological history, theoretical and semi-empirical relationships, and field measurements. Horizontal stresses can be measured in wells, and particularly in area of uplift and erosion they may exceed the vertical stresses. At high pore pressures this will result in horizontal fractures rather than vertical which is the case when the vertical stress is highest.

11.3.1 Theoretical and Semi-empirical Relationships

In a basin which is wide compared to its thickness, the compaction process due to added overburden may be modelled as a one-dimensional deformation situation (i.e. only strain in the vertical direction, no strain horizontally). This is often denoted as a uniaxial strain compaction situation. If the sediment mineral skeleton (framework) may be assumed to behave in a linearly elastic and isotropic manner (see Sect. 11.4), the horizontal stress which is built up as the vertical overburden is increased, is defined by:

$$\sigma'_H = \frac{\nu}{1-\nu}\,\sigma'_v \tag{11.5}$$

where ν is the Poisson's ratio for the sediment mineral skeleton. The same relationship would hold for unloading if the material really exhibits linearly elastic behaviour. Furthermore, for isotropic material, the magnitude of horizontal stress would be the same in all directions. If one assumes anisotropic behaviour, the equations corresponding to Eq. (11.5) would be somewhat more complicated, and the horizontal stresses would be different in the different directions. The horizontal stress coefficient for a uniaxial deformation situation is commonly called K_0 in geomechanics. For an assumed Poisson's ratio $\nu = 1/3$, K_0 becomes 0.5 from Eq. (11.5).

As discussed in Sect. 11.4, linearly elastic behaviour may be an acceptable approximation for a cemented sediment (sedimentary rock) undergoing minor deformation. However, during the initial gradual build-up of loose sediments in a basin, it is not realistic to assume linear elastic behaviour of the sediment framework. Its behaviour is very non-linear and inelastic, undergoing mainly permanent deformation. In soil mechanics one uses the following semi-empirical relationship for normally consolidated (NC) sediments. It is based on idealised theoretical considerations and on laboratory and field measurements:

$$K_{0\mathrm{nc}} = 1 - \sin\varphi' \tag{11.6}$$

where φ' is the angle of shearing resistance (friction angle) used in the Mohr-Coulomb failure criterion expressed in terms of effective stresses, and it is assumed that there is no cementation (cohesion $c = 0$). The K_0 value defined this way refers to the ratio of effective horizontal and vertical stresses (not total stresses). For $\varphi' = 30,\ K_{0\mathrm{nc}}$ becomes 1/2.

When such a sediment is unloaded (erosion) and thus becomes overconsolidated (OC), the horizontal stress does not decrease proportionally with the reduction in vertical effective stress because the sediment skeleton does not exhibit elastic behaviour. Horizontal stresses are "locked in" in the sediment, and the K_0-value increases. The following semi-empirical relationship is used, based on laboratory experiments and field measurements:

$$K_{0\mathrm{oc}} = K_{0\mathrm{nc}}(\mathrm{OCR})^n \tag{11.7}$$

where OCR is the overconsolidation ratio and n is a coefficient experimentally determined to usually be between 0.6 and 0.8, depending on the sediment properties. K_0 may well reach values above 1 (2–3 have been measured), which means that the horizontal stress is significantly larger than the vertical stress.

During burial and compaction and erosion (uplift), the horizontal stresses may change due to tectonic movements, and with time due to combinations of mechanical loading and unloading and also chemical compaction which may be independent of stress.

If extension occurs in a sediment with friction angle (φ') but no cohesion intercept (c), the effective horizontal stress coefficient would decrease from K_0 to a lower limiting value of:

$$K_{\mathrm{ext}} = 1 - \sin\varphi'/1 + \sin\varphi' \tag{11.8}$$

At this low lateral effective stress, shear failure would occur and a shear plane (normal fault) would form. Such extension may occur due to general basin extension, or more locally over a salt dome, over an elevated fault block of sedimentary rock with softer sediments on either side, or at the top of a slope.

If on the other hand, lateral compression occurs in the same sediment, the lateral effective stress coefficient would increase to a limiting value (reverse faulting):

$$K_{\mathrm{com}} = 1 + \sin\varphi'/1 - \sin\varphi' \tag{11.9}$$

Assuming $\varphi'= 30$, K_{ext} and K_{com} would be 1/3 and 3, respectively. If a cohesion intercept (c) is included, the value for K_{ext}would be smaller and K_{com} higher than given by Eqs. (11.8) and (11.9), respectively. For the case of horizontal compression the maximum horizontal effective stress is:

$$\sigma'_{\mathrm{H}} = \frac{1+\sin\varphi'}{1-\sin\varphi'}\,\sigma'_{\mathrm{v}} + 2c'\sqrt{\frac{1+\sin\varphi'}{1-\sin\varphi'}} \tag{11.10}$$

This is the linear Mohr-Coulomb failure criterion expressed in terms of the major and minor principal stresses. In this case σ_{H} is the major principal stress and σ_{v} the minor principal stress.

Zoback et al. (1985) and other investigators have measured high horizontal stresses in basement rocks. These stresses probably reflect compressional plate tectonic movements. However, basin sediments are much more compressible than the basement rocks (uncemented sands have been found as deep as 1.5–2 km in the North Sea). Therefore, the external plate tectonic and regional tectonic stresses that are transmitted through the underlying basement and the deeper well-cemented sedimentary rocks will have little effect on the horizontal stresses in the compressible sedimentary basin above, unless the lateral tectonic movements (compressive strains) are very large (Bjørlykke and Høeg 1997, Bjørlykke et al. 2005, 2006).

In the North Sea the horizontal stress has been found to increase with depth faster than the vertical stress, and at 4 km and deeper the total horizontal stress is usually 0.8–0.95 of the total vertical stress, approaching unity. The magnitude of horizontal stress at these depths is influenced by the effects of chemical compaction and creep. It should be noted that the ratio between the total horizontal and vertical stresses is not the same as the ratio between the effective stresses at the same location. In a sediment with high pore pressures (overpressure), and a ratio between total stresses of 0.9, the corresponding ratio between effective stresses may be about half that value, depending on the magnitude of overpressure. It is the ratio between effective stresses that indicates how close the sediment may be to a local shear failure, and it is the magnitude of the minimum effective stress that governs whether a tension fracture may occur.

The strain rates are important for the laboratory determination of K_0 because deformation by creep is a function of time. Some rocks behave very differently at low strain rates and at high strain rates. For further discussion on brittle and ductile behaviour see Sect. 11.4.4.

When the effective stress in the reservoir is increased due to reduced fluid pressure during petroleum production, the strain rates are fairly high. Therefore, the ratio between the horizontal and vertical stress will be controlled by mechanical compaction, and the K_0 values determined in laboratory tests may be applied.

11.3.2 Field Measurements of Horizontal Stress

As it is difficult to predict the horizontal *in situ* stress condition in the basin, and as the horizontal stresses may be very different in two orthogonal directions, it is common to resort to field measurements. The maximum horizontal stress is termed σ_H while the minimum horizontal stress is called σ_h. If the vertical stress is the major principal stress σ_1, the two horizontal principal stresses are σ_2 and σ_3, respectively.

To measure these two horizontal stresses and their orientation one may use so-called hydrofracturing tests in a borehole (e.g. Goodman 1989, Fjaer et al. 2008). For a typical situation where the major principal stress in the sediment is vertical, a vertical radial fracture will open in the wall of a vertical borehole when the fluid pressure in the borehole is increased. This is because the fracture will be oriented perpendicular to the direction of lowest effective stress which is in the horizontal direction. If the maximum and minimum horizontal stresses are different, the tangential stresses around the borehole vary. When the fluid pressure is equal to the minimum effective stress plus the rock tensile strength (τ) at the most critical location around the borehole, i.e. the location with the smallest initial tangential compression stress, a fracture opens in the wall of the borehole. This fluid pressure level is called the fracture pressure. By lowering the fluid pressure after the crack has opened, and then increasing the fluid pressure again, one may determine the tensile strength of the rock as the difference between the fracture pressure during the first and second load cycles. The tensile strength of a sedimentary rock is only a small fraction of the compressive strength, and tensile strength can often be neglected. Thus the fracture from the first load cycle may be used to determine the horizontal stresses.

The best way to measure the fracture pressure during the drilling phase of exploration is to perform a series of "minifrac" tests. However, this is not normally done, and it has become common practice in the petroleum industry to perform a simpler measurement that is called a "leak-off" test (Fig. 11.3). This is a pressure test in the well which is closed using the blow-out preventer valves. After the string of drill casing is set and cemented in the well, a leak-off test is normally run after a few metres of hole are drilled below the drill shoe. Mud is pumped into the well through the string using the cement pump of the drill rig. Return flow is prevented by cementing the casing, and the mud pressure is recorded as a function of time and injected volume. Before any fracture is opened, little or no mud is leaked into the sediment formation and the pressure simply increases as a linear function of the volume of mud injected. When the first fracture

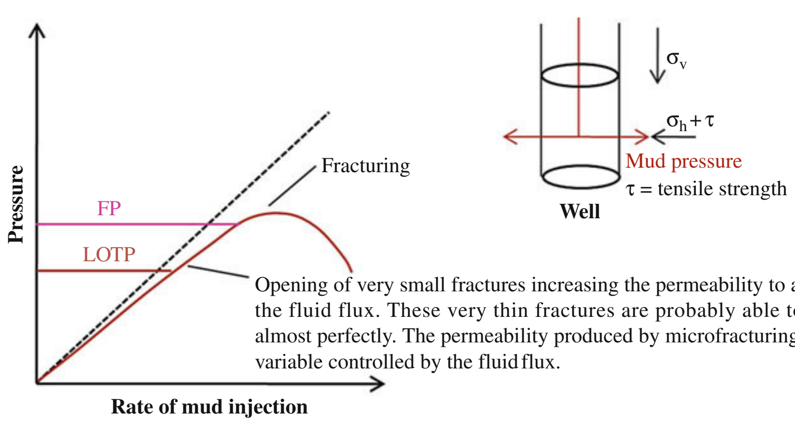

Fig. 11.3 Principle of leak-off test (LOT). The leak-off pressure must be higher than the lowest stress; fractures develop perpendicular to the direction of minor principal stress. In subsiding sedimentary basins the horizontal stress is usually lowest and the fractures will be vertical

is produced the pressure increase is reduced relative to the injected volume of mud, thus changing the slope of the curve. This leakage of mud into the formation is an indication that a thin fracture(s) has formed and that the fracture pressure has been reached (hence the name leak-off test). When repeated, the leak-off test fracturing starts a little earlier because now $\tau = 0$ across the fracture that was created during the first load cycle.

The results are often very reproducible which indicates that the rock was not seriously damaged during the first test, and that the small fractures produced probably closed again rather efficiently. The location of the fracture(s) may be determined by modern formation imaging tools called FMI. The location of the radial fracture gives the direction of the minimum principal stress as the latter is normal to the fracture (tangential to borehole wall at that location). The small fractures produced during a leak-off test may resemble the fractures formed during natural hydrofracturing at the top of an overpressure compartment in a sedimentary basin.

The directions of the maximum and minimum horizontal stresses may also be determined from borehole break-outs as shown in Fig. 11.4. The maximum compressive stress in the wall of the borehole occurs at each end of a diameter normal to the maximum horizontal stress direction. If the tangential stress is high enough to cause a compressive failure, a break-out occurs. By locating the break-out, one can determine the direction of the maximum horizontal stress. In practice the location may be determined by calliper measurements. The calliper measures the width of the borehole by recognising an oval shape. Modern formation imaging tools (FMI) can also be used to see the borehole break-outs.

11.4 Deformation Properties of Sedimentary Rocks

The fluid in the pores of the sediment compresses, but if there is no gas in the pore fluid, this effect is very small and insignificant when computing volumetric compaction in sedimentary basins. It is the compression bulk modulus of the grain structure that will govern the volumetric deformations, and the permeability and neighbouring drainage boundary conditions that will govern how quickly the fluid

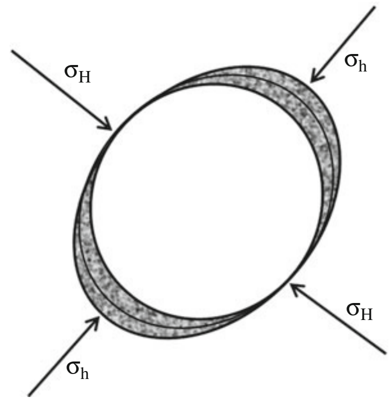

Fig. 11.4 Borehole break-out and the orientation of principal stresses. σ_H is the highest horizontal stress and σ_h is the lowest horizontal stress

may escape from the pores and allow the volume change to occur.

11.4.1 Concepts from the Theory of Elasticity

Elastic material behaviour, linear or non-linear, means that all strains (volume change and shear distortion) caused by a stress change are recovered when the stresses return to their original condition. If the grain skeleton (framework) of a sedimentary rock may be considered linearly elastic and isotropic for very small deformations, the deformational characteristics may be defined by the theory of elasticity.

Youngs modulus (E) is the ratio between the increase in normal stress and the resulting strain in the stress direction, when there is no change in the orthogonal normal stresses:

$$E = \sigma_z / \varepsilon_z \tag{11.11}$$

where σ_z is the applied stress and ε_z is the compressive strain in the z-direction, and $\Delta\sigma_x = \Delta\sigma_y = 0$. Poisson's ratio ($\nu$) is defined as $\varepsilon_x / \varepsilon_z$ in this situation. $_x$ is the strain in the x-direction and is equal to $_y$ if the material is isotropic. For a linearly elastic, isotropic material, only two constants (for instance E and ν) are required to fully define all the deformation characteristics.

The bulk modulus (K) is defined as the ratio between the increase in equal-all-round stress and the resulting volumetric compression (compaction):

$$K = \Delta\sigma/\varepsilon_{\mathrm{vol}} \tag{11.12}$$

For an isotropic material the bulk modulus is $K = E/3(1-2v)$. The shear modulus (G) is the ratio between the increase in shear stress and the resulting shear strain (angular change due to deformation). For an isotropic material it may be shown that $G = E/2(1+v)$.

For a uniaxial strain compaction situation with strain only in the vertical direction and no lateral strain allowed, the compaction (compression) modulus (M) may be expressed as $M = E(1-v)/(1+v)(1-2v)$. As stated in Sect. 11.3.1, the compaction process in a fairly homogeneous and wide sedimentary basin may be considered uniaxial. This is not the case in a narrow basin or where there are abrupt changes in depth of basin, lithology and material compressibility, for instance due to faulting and block rotations.

The strains induced by the transmission of seismic signals are so small that linear elastic behaviour may be assumed. However, anisotropic deformation characteristics should be allowed for when the seismic signals are used to derive deformation characteristics of the sedimentary rocks. A special type of anisotropy, which is a useful extension of the isotropic material theory, assumes the sedimentary rock to be transversely isotropic. This implies that the elastic properties are equal for all directions within a plane, but different in the other directions. Transverse isotropy may be considered to be a representative symmetry for horizontally layered sedimentary rocks. The properties are assumed isotropic, but different, in the vertical and horizontal planes. Five elastic constants fully define all the deformation characteristics for such a material.

In general, sedimentary rocks cannot be treated as linearly elastic materials because the normal and shear strains are not recovered when the element is unloaded. However, the modulus concepts from elasticity theory, as outlined above, are very useful even when analysing the more realistic non-linear behaviour of such rocks, including permanent (plastic) deformation.

It should be pointed out that for a linearly elastic ideal material like the one described above, there is no coupling between the effects of normal stresses and shear stresses and between normal strains and shear strains. For instance, if an element is only subjected to shear stresses, there will not be any normal strains. As discussed below, the behaviour of sediments and sedimentary rocks is not that simple, and there can be significant volume expansion (dilatancy) or contraction even if only shear stresses are applied. This phenomenon is also related to the degree of overconsolidation, as the higher the overconsolidation ratio, the more significant is the degree of shear dilatancy.

11.4.2 Non-linear, Inelastic Behaviour in Uniaxial Strain Compression

The stress–strain results from a saturated sediment tested in uniaxial strain compression are shown in Fig. 11.5a. The starting point (I) for the curve represents the initial state of stress, and it is assumed that initially the specimen is normally consolidated. The uniaxial compression modulus increases with the level of applied vertical effective stress. We therefore use the term tangent modulus (M_t) and/or secant modulus (M_s) to represent the behaviour. The tangent modulus gives the slope of the curve at any specified stress level, while the secant modulus gives the slope of the secant between two stress levels, commonly between the initial point I and point A. At stress level A the specimen is unloaded to the initial stress level (point B). As may be seen from the figure, a significant irrecoverable strain has been accumulated, given by the horizontal distance IB. The specimen is then reloaded back to A and up to a higher level, point C. It is found that the curve from A to C is a natural elongation of the curve from I to A. From I to A and A to C the specimen is normally consolidated, while during the unloading/reloading sequence it is overconsolidated. It should be noted that linearly elastic loading and unloading behaviour would be represented by only one common straight line in this diagram, from I to C.

Extensive laboratory testing of different types of sediments has shown that the tangent modulus may be determined by the following general expression:

$$M_t = m\,p_0\left(\frac{\sigma'_v}{p_0}\right)^{1-a} \tag{11.13}$$

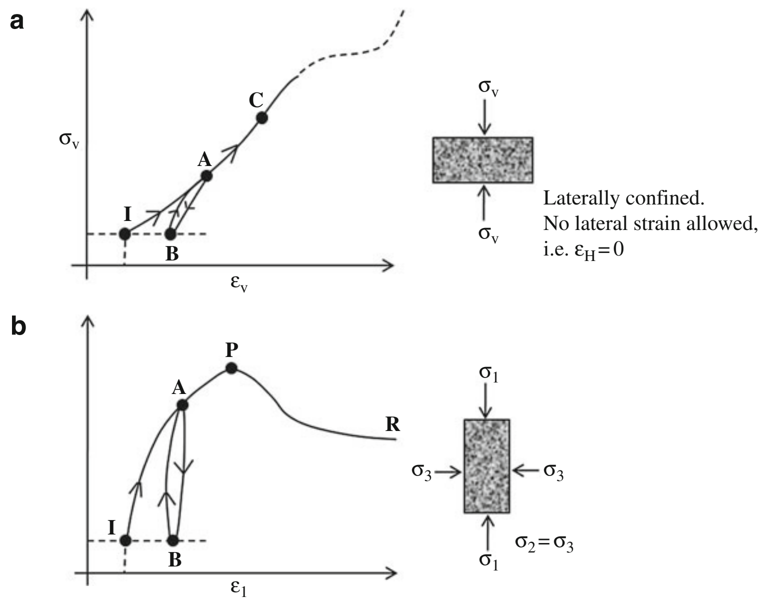

Fig. 11.5 Stress–strain behaviour of sedimentary rocks. (**a**) Stress–strain behaviour of sediment during uniaxial loading, unloading and reloading. (**b**) Stress–strain behaviour of sediment under triaxial compression with no restraint of lateral strains; P is the peak strength and R is the residual strength

where p_0 is a reference stress to make the ratio inside the parenthesis non-dimensional (often p_0 is set equal to 0.1 MPa in the published literature), and "m" and "a" are non-dimensional coefficients depending on the type of sediment and its geological history. To fit the different experimental curves, the exponent "a" is found to lie between 0 and 1. For a normally consolidated clay sediment $a = 1$ and this fits quite well, and the effective stress $\sigma' v$ gives an M_t value which is directly proportional to σ_v. For overconsolidated clay sediments (weak claystones) an a-value of 0 gives a fair approximation, and that corresponds to an M_t which is constant as given by the almost straight lines for the unloading and reloading stress–strain curves in Fig. 11.5a.

When the effective vertical stress is increased much beyond the C-level, one may find that the stress–strain curve bends rather sharply over to the right (see dotted curve in Fig. 11.5a) before it again starts to rise for a further increase in the vertical effective stress. In a sand this is caused by crushing of the coarser sand grains because the intergranular contact stresses become so high (Chuhan et al. 2002, 2003); this is further discussed in Sect. 11.5. A similar phenomenon happens in sediments with a cemented but open and porous structure. This was clearly demonstrated for the reservoir chalk at the Ekofisk Field in the North Sea. When the effective vertical stress was increased by reducing the fluid pressure in the reservoir, it reached a level at which the coccolith structure (framework) of the chalk collapsed and caused large vertical strains, reservoir compaction and seafloor subsidence. Subsequent injection of seawater maintained the fluid pressure in the Ekofisk reservoir and avoided further increase of effective stresses. The rate of compaction was reduced, but not stopped, because it was later found that the seawater weakened the chalk framework and increased the compressibility.

11.4.3 Non-linear, Inelastic Behaviour When Approaching Shear Failure

An element in a state of true uniaxial strain compression (Sect. 11.4.2) can undergo large vertical compression (compaction), but it cannot fail in shear by creating failure planes and fractures. This is prevented by the lateral confinement of the element (one-dimensional compression).

However, consider now a cylindrical specimen of a saturated sediment/sedimentary rock as shown in Fig. 11.5b. The specimen is first subjected to an axial stress (σ_1) and a radial stress (σ_3) representing the initial *in situ* stress condition. The initial pore pressure in the specimen is also the same as the one *in situ* and thus the initial effective stresses. Then the axial stress is increased while the lateral stress is kept constant. The loading is performed so slowly that any tendency to overpressure in the pore fluid is avoided by allowing the fluid to drain out of the specimen (dissipate), so that the pore pressure remains at its initial value. In geomechanics this is called a drained test, as compared to an undrained test in which the fluid is not allowed to drain and overpressures (positive or negative) build-up during the loading.

The recorded stress–strain curve for this axial loading is shown in Fig. 11.5b. Loading occurs from the initial point I to point A. The initial section of the curve is fairly straight (linear), but as the axial stress increases the curve starts to bend. The slope of the curve at any point is called the tangent Young's modulus (E_t). The secant from point I to A gives the secant modulus (E_s) up to that stress level. If the axial stress at point A is reduced down to the original axial stress level, the unloading curve goes down to point B and an irrecoverable strain given by the distance IB has been accumulated. Upon reloading the curve climbs back up to point A. The slopes of the unloading-reloading curves are very similar and close to the initial slope of the curve I to A. When the axial stress is increased beyond point A, the curve continues to bend as the specimen approaches a shear failure condition. Both the tangent and secant modulus decrease significantly. As the stress difference ($\sigma_1 - \sigma_3$) becomes even larger, so does the maximum shear stress in the specimen. The stress difference cannot exceed a certain level (strength), as the specimen is not able to carry any more load, and large axial strains (and shear strains) ensue.

If one were to start the test described above with a higher horizontal effective stress level, the stress–strain curve would be steeper and climb higher. For loose sediments the moduli and shear strength are strong functions of the horizontal effective stress level, therefore it is so important to be able to estimate the effective stresses. For sedimentary rocks (shales and sandstones) with strong cementation caused by chemical processes, the modulus and strength are also influenced by the horizontal effective stress, though to a much smaller extent. Note that the triaxial test shown in Fig. 11.5b may also be run as a uniaxial strain test. This is done by adjusting the horizontal stress at all stages of the test such that no horizontal strain is allowed to occur. This is called a K_0 test.

11.4.4 Brittle Versus Ductile Stress–Strain Behaviour

For some sedimentary rocks, and depending on the magnitude of the lateral effective stress, the stress–strain curve in Fig. 11.5b may drop abruptly after it has reached a peak at point P (the peak strength). This is accompanied by a drop in shear resistance with further strain down to a level R which is called the residual strength after the failure. The behaviour from the peak down to residual is termed strain-softening or strain-weakening.

For other sedimentary rocks there is no strain-softening, and the stress–strain curve is fairly horizontal or even slightly climbing as the strain increases. This is called ductile behaviour. A sedimentary rock found to behave as a brittle material at low horizontal effective stress, may well behave as a ductile material when the effective horizontal stress becomes sufficiently high, i.e. the horizontal stress has reached the brittle-to-ductile transition stress level (e.g. Goodman 1989).

Other factors also affect the degree of brittleness. The higher the temperature and the lower the stain rate, the less tendency for brittle behaviour. In this connection it should be pointed out that the rate of stress change in and around a reservoir during petroleum production is orders of magnitude higher than the geological stress changes, except for earthquake occurrences. Some rocks behave very differently at low strain rates compared with at high strain rates. Rock salt is brittle when loaded at a very high strain rate, but flows like a viscous fluid over geological time. Also carbonate sandstones and shales yield to stress in a ductile manner by mechanical as well as chemical compaction if the strain rate is low enough.

11.5 Compaction in Sedimentary Basins

The deformation characteristics for sedimentary rocks are complex as shown above. The strain rate in sedimentary rocks is normally rather low except during tectonic deformation like faulting. During production of reservoir rocks the stain rates may become high.

In sedimentary basins down to depths of 1.5–2 km mechanical compaction caused by the increase of effective vertical stresses is the dominating compaction process and has the greatest influence on the sediment's hydro-mechanical properties. However, at greater depths where temperatures are higher (>70–80°C), it is mainly chemical compaction that contributes to the volume change and to the hydro-mechanical properties through the effects of dissolution, precipitation and cementation.

11.5.1 Sands and Sandstones

The effective stress of the overburden is transmitted through a framework of load-bearing grains. These grain-to-grain contact stresses may become very much higher than the average effective stress. Not all grains will be subjected to high stresses because they may be shielded by other grains within the load-bearing grain framework. The stress from the overburden will thus be concentrated on these other grains, which then may fracture. The small areas of grain contact make the stress at these contacts very high, even at moderate burial depth.

Natural sand grains of quartz and feldspar are mostly blocky rather than spherical and have an irregular surface. This means that the area of contact is likely to be very small even for the larger grains, resulting in much higher contact stresses than for smaller grains. If sand grains had been perfectly spherical the contact would be controlled by the elasticity, and the stress would then be independent of the grain size.

Experimental compaction of well sorted sand reveals that coarse-grained sand aggregates are subjected to significant grain fracturing and compaction at 20–30 MPa effective stresses while fine-grained sand does not fracture, and compacts much less at the same stress levels (Fig. 11.6) (Chuhan et al. 2002, 2003, Bjørlykke et al. 2004).

Well sorted sand (e.g. beach sand) with an intitial porosity of 40–45% may compact mechanically to 30–38% porosity, depending on the stress level and the grain size. Poorly sorted sand and sand with high mud contents will compact much more at even lower stresses.

With increasing compaction due to grain rearrangement or breakage, and cementation, the rock becomes less porous, less compressible and stronger. This process increases both the number and area of grain contacts.

Loose uncemented sands subjected to shear deformation may develop thin deformation bands. Such shear bands may be composed of densely packed grains where smaller silt-sized grains have been packed between larger grains during the shearing. If there is little clay the shear strength of the shear band will exceed the shear strength of the matrix and the shear deformation will shift laterally to an area where there has been no strain (deformation). This is called strain-hardening and results in a network of shear bands which have only been subjected to small offsets. Large displacements recorded on seismics may in reality consist of a broad zone of deformation bands.

Shear deformation in sand may also result in grain crushing. Experimental deformation of sand suggests, however, that the effective stress normal to the shear band must be at least 10 MPa for coarse sand to be crushed. In the case of normal faults the horizontal stress must have been 10 MPa and the vertical stress 20–25% higher, which corresponds to burial depths of 1–1.5 km.

Precipitation of quartz or other cements increases the stiffness of sand and reduces its compressibility, transforming loose sand into indurated sandstones. Only relatively small amounts of quartz cement, probably only 2–4%, are required to effectively stop the mechanical compaction that is due to rearrangement of grains. As a result the velocity, and particularly the shear velocity, will increase sharply for a modest reduction in porosity.

Sandstones may behave as if they were overconsolidated due to cementation (chemical compaction) and will only compact following the stress-strain curve for overconsolidated rocks. We must distinguish between overconsolidation due to previously higher effective stresses and "pseudo overconsolidation" caused by cementation and chemical compaction. This because the highest effective stress must be estimated from the burial curve and the pore pressure while the chemical compaction may be rather insensitive to changes in stress.

Chemical compaction involving the dissolution and precipitation of quartz is controlled mainly by temperature because the rate of quartz cementation seems to be the rate-limiting step rather than the effective stress. The mineral cement growing as overgrowth on detrital quartz grains into the pore space between the grains is not subjected to stress at the time of cement precipitation and is not influenced by the stress. However, after further compaction it may become a part of the load-bearing structure. Therefore the grain-to-grain contact stresses in a sandstone may be lower at depths of 3–4 km than at 1–2 km because the stress is distributed over a larger grain contact area and is also supported by quartz cement. Most mechanical grain crushing therefore occurs at depths shallower than about 2–3 km where there is little quartz cement present. In the case where quartz cementation (overgrowth) is prevented by coatings (e.g. chlorite), grain fracturing may occur at 3–4 km at about 35–40 MPa effective stresses. Grain fracturing will then expose uncoated fresh quartz surfaces for quartz cementation which gradually will reduce the porosity (Fig. 11.6).

During production of a reservoir, reduced pore pressure and higher effective stress will cause compaction of the reservoir rocks. In the case of well-cemented reservoir sandstones this effect is very small, but in shallow reservoirs with loose sand such compaction can be significant.

11.5.2 *Compaction of fine grained sediments*

11.5.2.1 Clays and Mudstones

Clays, mudstones and shales have very different physical properties compared to coarser-grained sediments like siltstones and sandstones. The physical properties of clays depend not only on the strength of the sediment particles (mostly clay minerals), but also on their surface properties and chemical bonds which are controlled by the composition of the pore fluids. Clays,

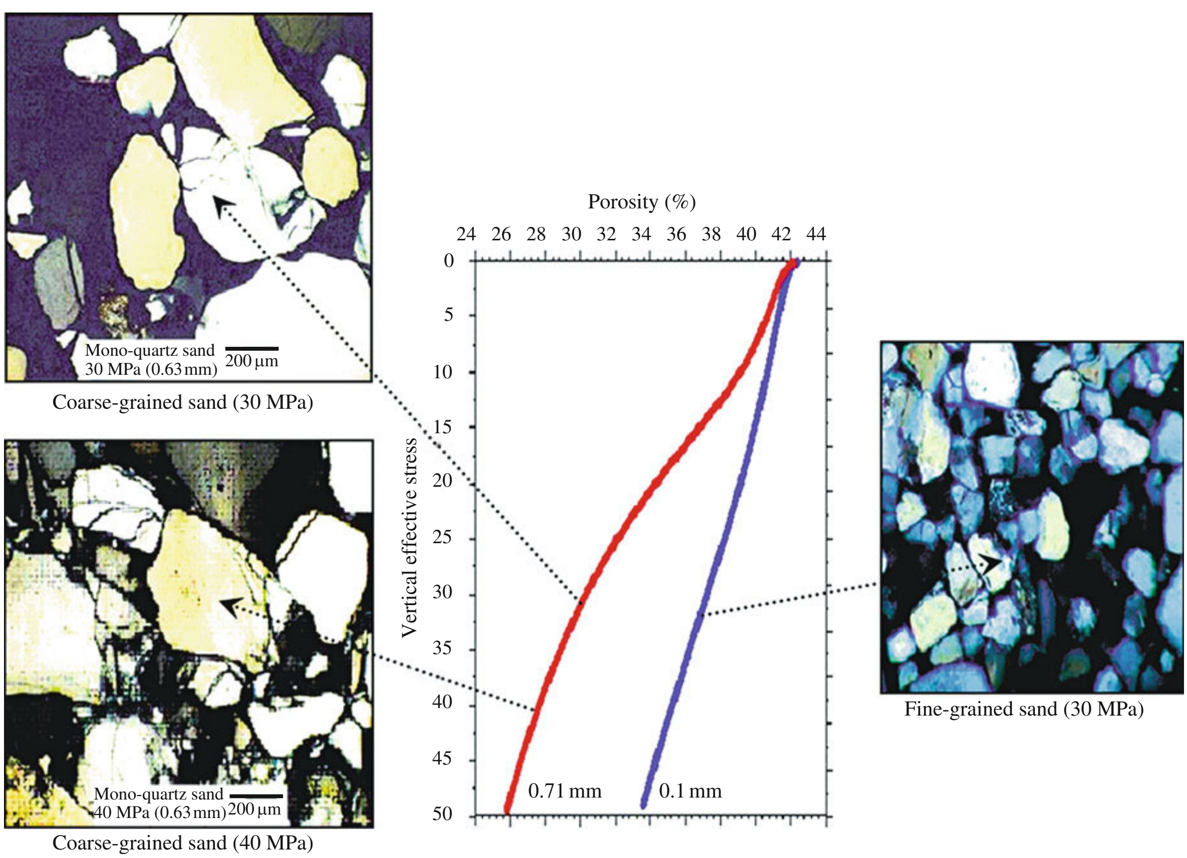

Fig. 11.6 Experimental compaction of loose sand grains (after Chuhan et al. 2002). Coarse-grained sand is more compressible than fine-grained sand. This is because there are fewer grain contacts and more stress per grain contact in coarse-grained sand, resulting in more grain fracturing

mudrocks and shales may vary greatly and have very different physical properties, depending on the clay mineral composition and on the content of silt and sand. Mudstones and shales may have high contents of silt and sand and also carbonate and may from transitions into poorly sorted sandstones and impure limestones.

The most common clay minerals are smectite, illite, chlorite and kaolinite. When the silt and sand content exceeds 40–50% there may be a grain-supported structure, and this marks the transition into clay-rich siltstones and sandstones. A sediment of sand and silt grains floating in a matrix of clay has for many purposes the same properties as the clay fraction, but the density is higher and so is the seismic velocity. This is because the stiffness is determined by the load-bearing grains.

Poorly sorted clays like glacial clays compact readily at relatively low effective stresses because the clay particles are packed in between the silt and sand grains. Data on compaction of clay and mudstones may be obtained from measuring the degree of compaction where the burial history and maximum effective stress can be estimated, or by experimental compaction in the laboratory.

In fine-grained sediments like clays and mudstones the total overburden stress is distributed over a very large number of grain contacts and the stress per grain contact may be quite low. Relatively small amounts of minerals precipitated as cement between the primary grains can then cause a very significant increase in stiffness and seismic velocity even at shallow burial. Most commonly this involves carbonate cement, which makes soft clay grade into marls and calcareous mudstones with higher bulk moduli. Quartz cementation requires higher temperatures (>80°C) corresponding to 2–2.5 km in basins with normal geothermal gradients.

11.5.2.2 Experimental Compaction of Clays

Experimental compaction of clays is difficult. It requires very careful sample preparation, and compaction tests up to 50 MPa stress may take 5–6 weeks for smectite-rich clays. This is because the permeability is so low that it takes a long time for the excess water to drain. Time is also required to allow for the slight compaction at constant stress (creep) which may also be referred to as secondary compaction.

Compaction of mud to mudstones and shales is the result of natural processes during burial, usually over several million years. We can determine the resultant rock properties by analysing natural rock samples in the laboratory. It is nevertheless still difficult to estimate the effective stress and temperatures to which these rocks have been subjected. This is particularly true in the case of samples exposed on land after substantial uplift. Samples from offshore wells in subsiding basins are much better constrained with respect to the burial history but representative mudstones are rarely cored. Cuttings can be analysed mineralogically, but it is difficult to test their mechanical properties without reconstituting the samples.

There is often a need to predict the compaction of sediments (soils) including clays in an engineering context and they are then tested in the laboratory to measure the strain (compaction) as a function of effective stress and other soil and rock mechanical parameters. To simulate natural burial in sedimentary basins we use rather high stresses, up to 50 MPa or more, corresponding to 4–5 km of overburden. In most cases, though, chemical compaction becomes dominant at shallower depth.

By testing artificial mixtures of clays we can measure their physical properties as a function of clay mineralogy and silt and sand content. Kaolinitic clays compact much more readily than smectite, which is the most fine-grained clay mineral and has very low compressibility (Mondol et al. 2007). At about 20 MPa effective stress, corresponding to about 2 km of burial, pure smectite has more than 40% porosity while kaolinite has less than 20% (Fig. 11.7). Even at 50 MPa corresponding to 4–5 km burial depth at hydrostatic pressure the porosity may still exceed 40%. Clay minerals compact more when wet than dry (Mondol et al. 2007). This is probably because the friction between the grains is higher in dry clays (Fig. 11.7).

In the case of smectite the large surface area and the water which is bound to these clay surfaces make it difficult to define the proportion of free water, and hence determine the exact porosity, which will also depend on the composition (electrolytic strength) of the porewater. Clay minerals tend to have a negative charge, causing repulsion between the clay particles. However, the negative charges will adsorb cations like Na^+ and K^+, thus neutralising this repulsion. This is what causes flocculation when river-borne clays enter the sea. Marine clays therefore have a more stable clay mineral fabric and higher shear strength than

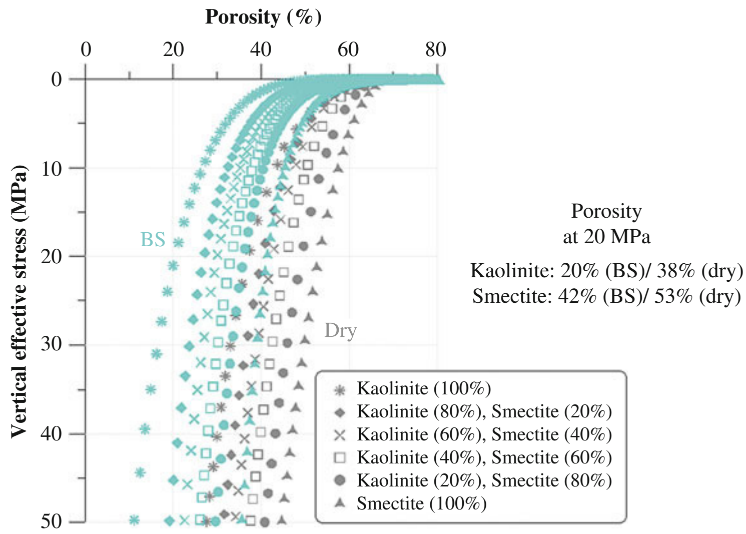

Fig. 11.7 Experimental mechanical compaction of dry (in *grey*) and brine-saturated (in *colour*) clay aggregates under uniaxial compression strain (after Mondol et al. 2007). Porosity at 20 MPa effective stress of dry and brine-saturated pure smectite and kaolinite mixtures is shown

freshwater clays. Slow weathering and leaching of saltwater (Na^+, K^+) from marine clays uplifted by glacial unloading in Scandinavia may therefore lead to slope instability and "quick clay" slides. Experimental compaction of clays confirms that their compressibility and shear strength are a function of the salinity of the porewater.

Smectitic clays are less compressible than kaolinitic clays and this may be partly due to chemical bonds and partly because of the fine grain size. The negative charge of the clay minerals also causes water to be bound to the mineral surfaces by the positive charge of the dipole of the water molecule. This is important in the case of smectite which has a surface area of several hundred m^2/g but is not so significant for coarser clay minerals like illite, chlorite and kaolinite which have much lower surface area bound water. The total stress is divided by the number of grain contacts and the stress per grain is then less in smectitic clays. It has been shown experimentally (Mondol et al. 2008a) that fine-grained kaolinite is less compressible than coarse-grained kaolinite (Fig. 11.8). Similarly, well sorted fine-grained sand is less compressible than coarse-grained sand (Chuhan et al. 2003). Smectitic clays are characterised by low velocities and low density because they have high porosity (Fig. 11.9) (Mondol et al. 2008b).

11.5.2.3 Chemical Compaction of Clays and Mudstones

Clays compact mechanically at shallow depth and at temperatures below 70–80°C, but at higher temperatures compaction may be controlled by chemical reactions. Chemical compaction must have thermodynamic drive so that less stable minerals dissolve and more stable minerals precipitate. Clay minerals like smectite become unstable and are replaced by mixed-layer minerals and illite. The silica released by this process must be precipitated as quartz cement for this reaction to proceed and this causes marked stiffening and higher velocities. Compaction is then mainly controlled by temperature rather than effective stress.

Amorphous silica from volcanic sediments, and from amorphous silica (opal A) from fossils like diatoms and siliceous sponges, will be a silica source for precipitation of quartz cement even at low temperature. Carbonate cements from calcareous fossils will also result in a marked increase in the stiffness and velocity. In the absence of thermodynamically unstable minerals like smectite, mudstones may remain nearly uncemented to greater depth. At about 130°C kaolinite becomes unstable in the presence of K-feldspar and causes precipitation of illite and quartz. Gradually, however, mudstones become harder and develop a schistocity, becoming a shale. The cleavage

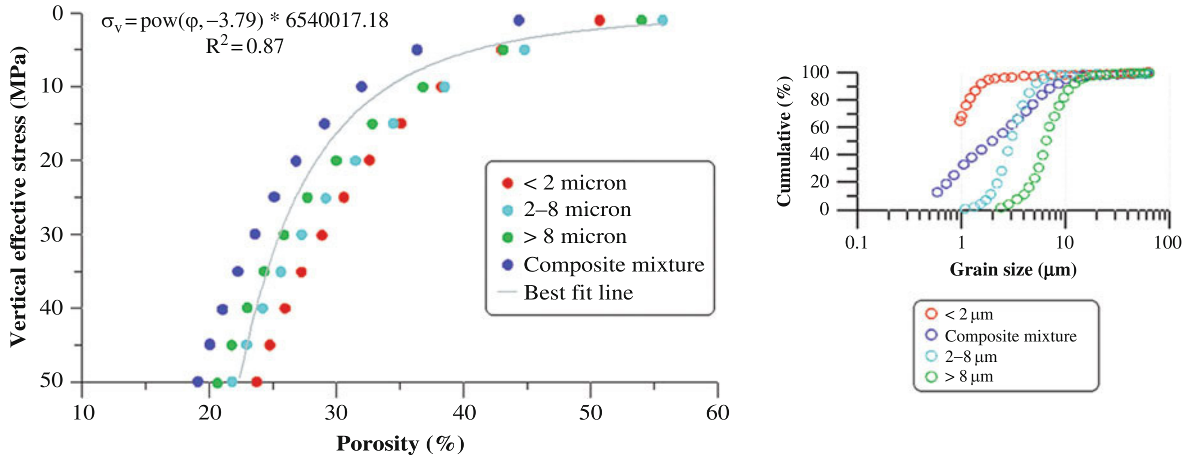

Fig. 11.8 Experimental mechanical compaction of brine-saturated kaolinite aggregates sorted by grain size (after Mondol et al. 2008). The sample containing less than 2 µm sized kaolinite aggregates retained higher porosity compared to all the other mixtures. The maximum porosity reduction is observed in the composite mixture containing all the grain sizes, demonstrating the importance of both grain size and sorting for the rock properties

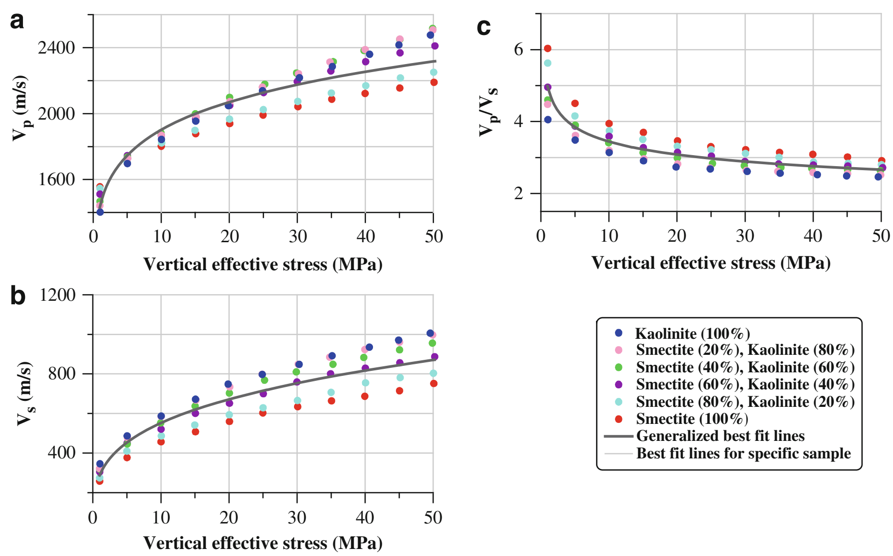

Fig. 11.9 Crossplots of ultrasonic velocities vs. vertical effective stress of brine-saturated smectite-kaolinite (modified after Mondol et al. 2008b). Pure smectite has lower V_p (**a**) and V_s (**b**) compared to pure kaolinite. The V_p/V_s ratio is higher in pure smectite than in pure kaolinite (**c**)

is produced by pressure solution of quartz and other minerals in contact with layers enriched in clay minerals.

The transition from mudstones to shales does not only involve a marked increase in stiffness and velocity but also an increase in the anisotropy. This is due to the reorientation of clay minerals during diagenesis so that the velocity parallel to bedding will be higher than perpendicular to bedding. In addition the resistivity will to a large extent be controlled by the orientation of the clay minerals and the quartz cementation in mudrocks and shales.

11.6 Summary

In subsiding sedimentary basins, mechanical compaction caused by the increase of effective vertical stresses is the dominating compaction process and has the greatest influence on the sediment's porosity and hydro-mechanical properties down to depths of 2–2.5 km. At greater depths where temperatures are higher (>70°C), it is mainly the chemical compaction that contributes to the volume change and to the hydro-mechanical properties due to the effects of dissolution, precipitation and cementation. Carbonate rocks, though, may undergo chemical compaction at shallower depths.

The magnitude and distribution of stresses in sedimentary basins are important in relation to petroleum exploration, production and reservoir management. Knowledge of *in situ* stress is also important in connection with drilling, particularly during deviation and horizontal drilling. The magnitude and orientation of stresses affect the propagation and interpretation of seismic signals, particularly the S-waves, through sedimentary rocks.

The total vertical stress (σ_v) of a rock sequence is carried partly by transmission of stress in the solid grain framework (effective stress σ'_v) and partly by the pressure in the fluid phase (porewater or petroleum).

Determination of effective stresses depends on reliable estimates of the fluid pore pressure which often is in excess of hydrostatic (overpressure). There exist semi-empirical relationships for the ratio between horizontal and vertical effective stresses, but reliable estimates of horizontal stresses depend on field measurements like hydraulic fracturing tests. It is common practice in the petroleum industry to use a simplified procedure (leak-off tests) to determine the magnitude and orientation of the minimum horizontal stress.

The virgin (*in situ*) distribution of stresses in sedimentary basins is the result of both mechanical and chemical compaction, usually over geological time. Changes in stresses during petroleum production from a reservoir are much more short term and are mainly mechanical. In a carbonate reservoir the chemical processes may be so fast that also chemical compaction may become significant at that time scale.

The effective stresses in sand may cause grain-to-grain contact stresses which are so large that compression (compaction) may occur due to crushing and fracturing of grains. This is more pronounced in coarse-grained rather than fine-grained sands and may account for a significant component of the porosity reduction. After the grain crushing and permanent collapse deformations have occurred, the grain size is reduced and the reservoir regains stiffness. With increasing depth such grain crushing is less likely due to increased cementation of the grain structure caused by chemical processes. In most sandstone reservoirs quartz cementation starting at 2–2.5 km (70–80°C) will stabilise the grain framework and prevent further mechanical compaction. Sandstone reservoirs with a critical content of quartz cement (>2–3%) will therefore experience very little compaction even if the effective stress is increased during production. Similar compaction by grain breakage may also occur for high effective stresses in a reservoir where the framework of the sedimentary rock is very porous with little cement (e.g. the chalk in the Ekofisk reservoir, North Sea). Here this led to very large reservoir compaction and subsequent seafloor subsidence (c. 10 m).

Smectitic clays are characterised by very high V_p/V_s ratios when compared with other clays (Fig. 11.9c). This implies that in a sequence of mudstones, smectitic clays will stand out with very different characteristics compared with kaolinite and probably also illite-rich sequences, which have much lower velocities and V_p/V_s ratios. At temperatures above 70–80°C smectite will no longer be stable and mudstones will be more influenced by chemical compaction and cementation. In cold basins, however, mechanical compaction can be dominant down to 4–5 km burial depth. Every mudstone has a unique compaction curve which will depend on a number of factors such as mineralogy, grain size, pore fluids, pore pressure, pore aspect ratio, etc.

Well log data from well sorted Jurassic sandstone like the Etive Fm in the North Sea basin show that natural compaction curves agree well with experimental mechanical compaction in the laboratory (Fig 4.16).

The physical properties of silt and clay mixtures depend strongly on both the clay mineralogy and the content of silt and sand. (Manzar et al. 2010). This includes porosity and velocity and also anisotrophy as measured in the laboratory. The properties of mudstones are also influenced by carbonate and silica cement.

The effect of time on mechanical compaction will always be difficult to evaluate in the laboratory but prolonged compaction tests suggest that compaction due to creep is rather small. At depths shallower than about 2 km, high porosity (>30%) is preserved in Upper Palaeozoic sandstones despite long burial time.

In spite of the limitations, laboratory porosity/density/velocity-stress relations and their comparison with data found in well logs will provide important constraints on evaluation of burial depths, pore pressure prediction and/or the amount of uplift and erosion found in sedimentary sequences (Mondol et al. 2007, 2008b, c,).

Further Reading

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