Chapter 11

Introduction to Geomechanics: Stress and Strain in Sedimentary Basins

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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:

$$\sigma_v = \rho_b g h \tag{11.1}$$

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):

$$\sigma'_v = \sigma_v - u \tag{11.2}$$

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:

$$\rho_b = \varphi \rho_f + (1 - \varphi)\rho_m \tag{11.3}$$

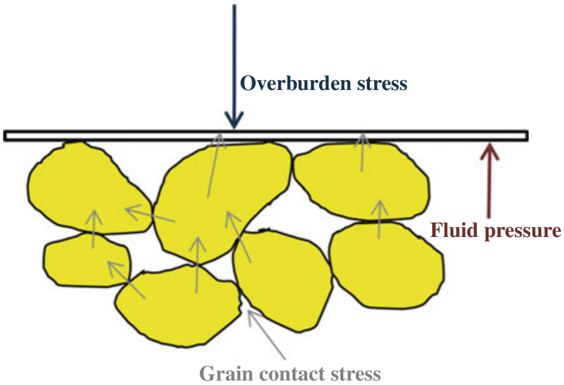

Fig. 11.1 The total vertical stress from the overburden (σ'_v) is carried by the mineral grain framework (solid phase) and the pore pressure (fluid phase). The effective stress is defined as the overburden vertical stress minus the pore pressure and it is transmitted through the grain contacts

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

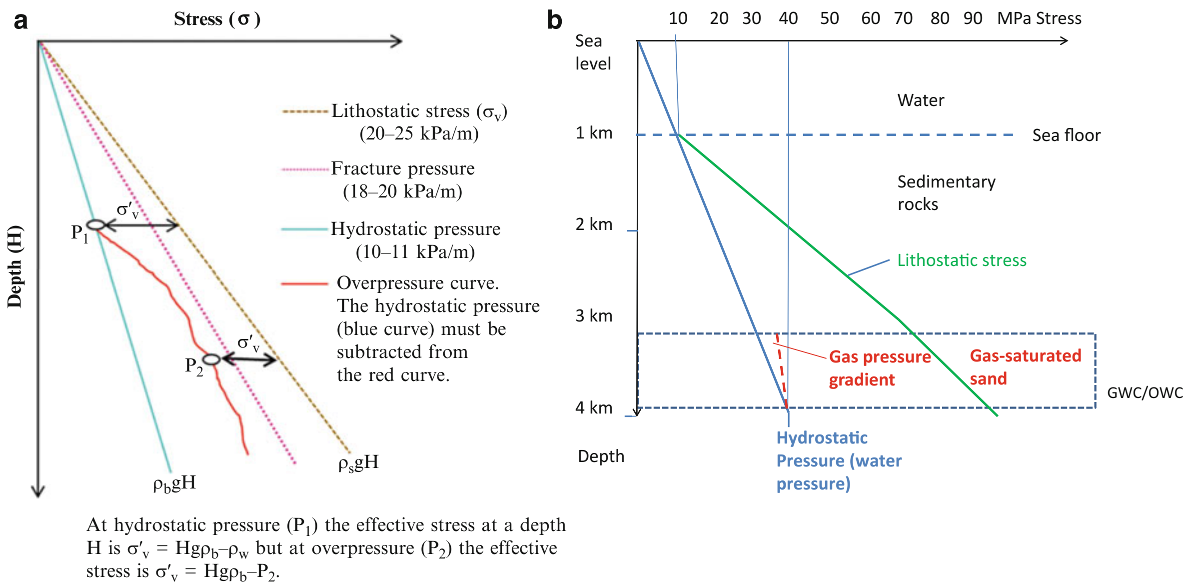

Fig. 11.2 (**a**) Simplified diagram showing the increase in vertical total stress (lithostatic) and hydrostatic pressure as a function of depth. In reality these lines are not strictly straight because the total vertical stress varies as a function of the sediment bulk density (ρ_b) which tends to increase 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:

$$\sigma'_H = \frac{\nu}{1-\nu}\,\sigma'_v \qquad (11.5)$$

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:

$$K_{0nc} = 1 - \sin\varphi' \qquad (11.6)$$

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 $\varphi' = 30,\ K_{0nc}$ 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:

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

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

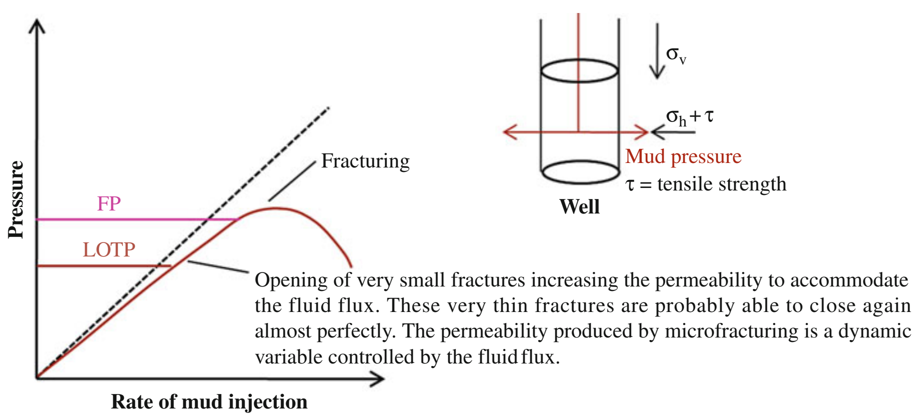

Fig. 11.3 Principle of leak-off test (LOT). The leak-off pressure must be higher than the lowest stress; fractures develop perpendicular to the direction of minor principal stress. In subsiding sedimentary basins the horizontal stress is usually lowest and the fractures will be vertical

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.

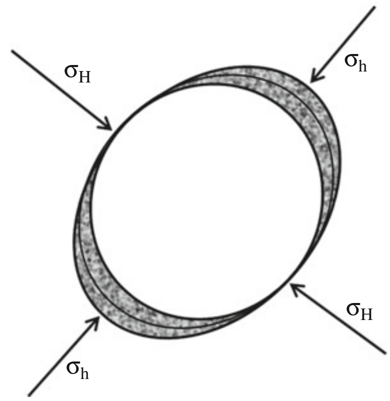

Fig. 11.4 Borehole break-out and the orientation of principal stresses. σ_H is the highest horizontal stress and σ_h is the lowest horizontal stress

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 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:

$$E = \sigma_z / \varepsilon_z \qquad (11.11)$$

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:

$$M_{\mathrm{t}} = m\,p_0 \left(\frac{\sigma'_v}{p_0}\right)^{1-a} \tag{11.13}$$

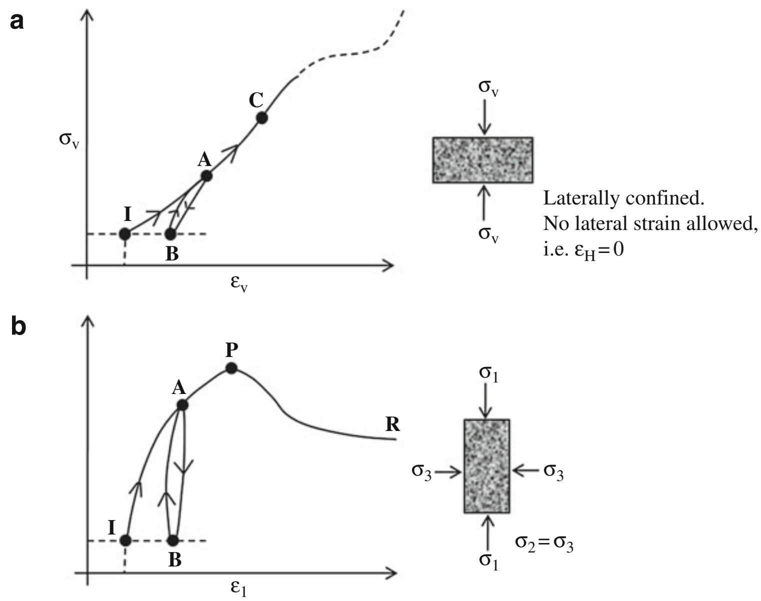

Fig. 11.5 Stress–strain behaviour of sedimentary rocks. (**a**) Stress–strain behaviour of sediment during uniaxial loading, unloading and reloading. (**b**) Stress–strain behaviour of sediment under triaxial compression with no restraint of lateral strains; P is the peak strength and R is the residual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

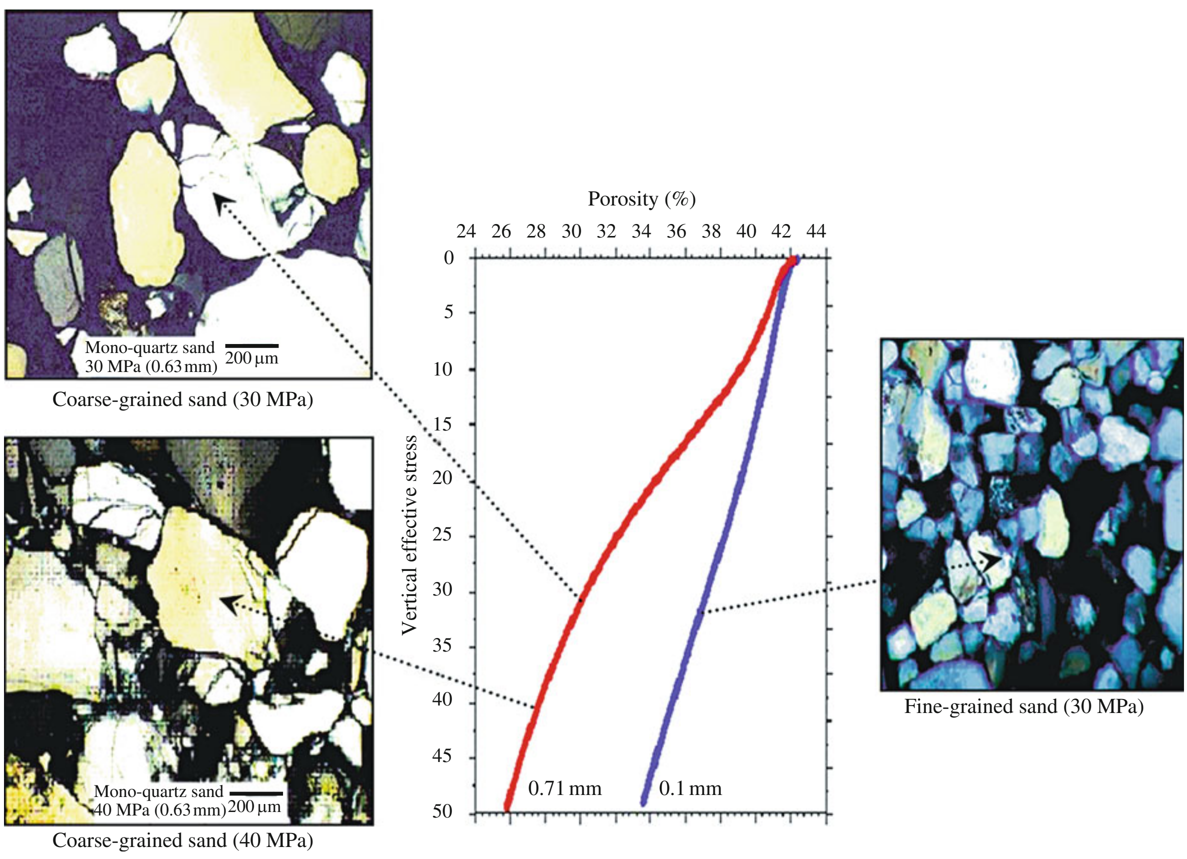

Fig. 11.6 Experimental compaction of loose sand grains (after Chuhan et al. 2002). Coarse-grained sand is more compressible than fine-grained sand. This is because there are fewer grain contacts and more stress per grain contact in coarse-grained sand, resulting in more grain fracturing

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

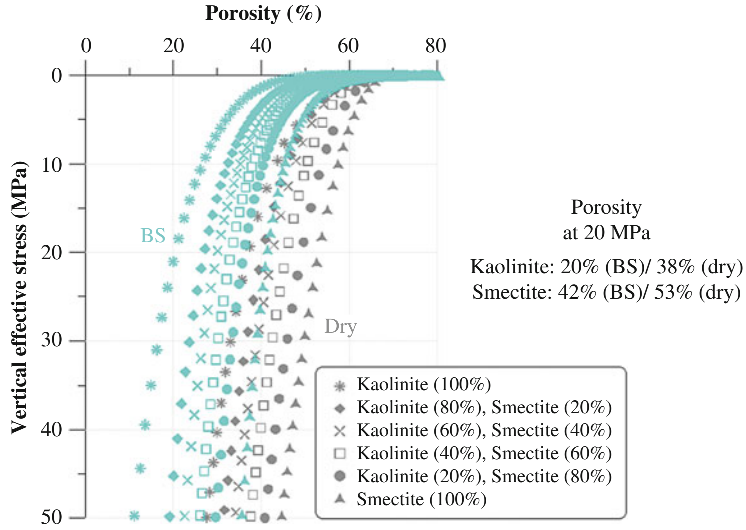

Fig. 11.7 Experimental mechanical compaction of dry (in *grey*) and brine-saturated (in *colour*) clay aggregates under uniaxial compression strain (after Mondol et al. 2007). Porosity at 20 MPa effective stress of dry and brine-saturated pure smectite and kaolinite mixtures is shown

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

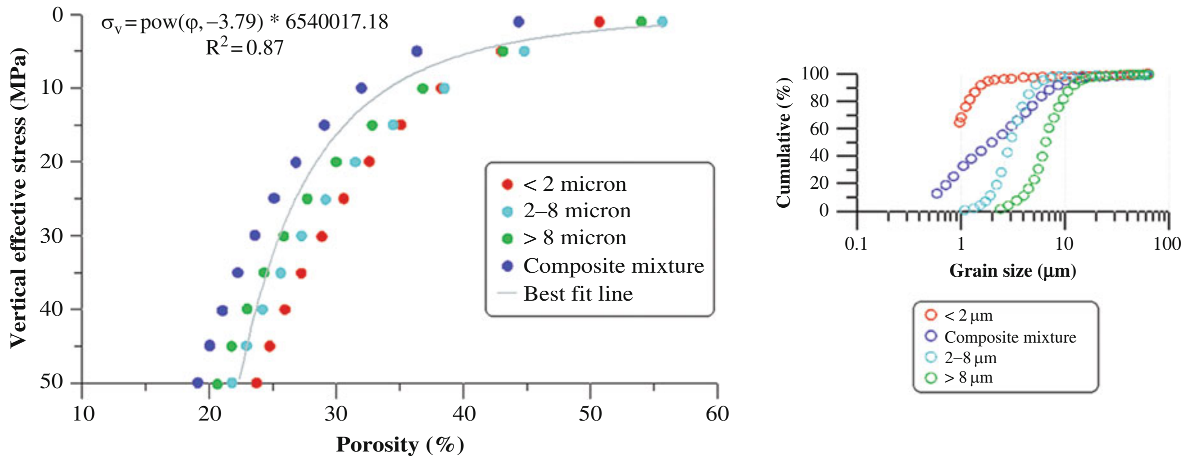

Fig. 11.8 Experimental mechanical compaction of brine-saturated kaolinite aggregates sorted by grain size (after Mondol et al. 2008). The sample containing less than 2 µm sized kaolinite aggregates retained higher porosity compared to all the other mixtures. The maximum porosity reduction is observed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

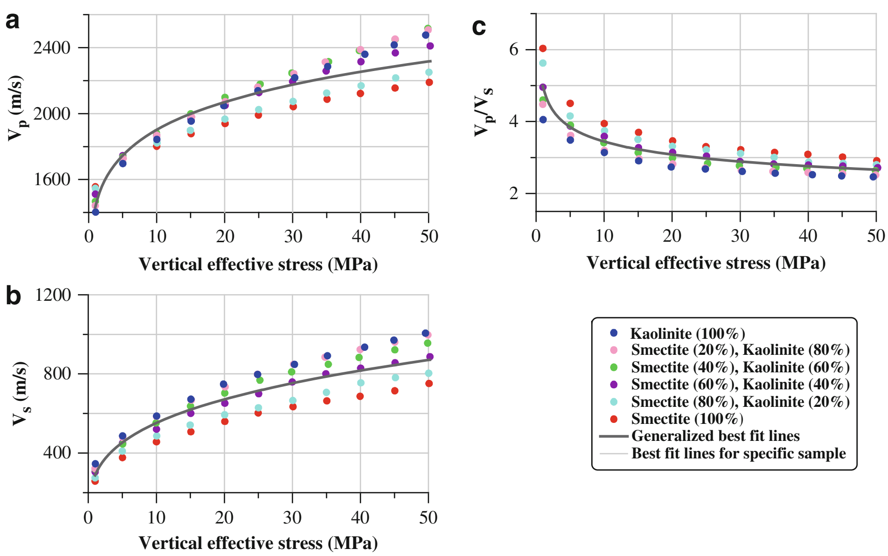

Fig. 11.9 Crossplots of ultrasonic velocities vs. vertical effective stress of brine-saturated smectite-kaolinite (modified after Mondol et al. 2008b). Pure smectite has lower V_p (**a**) and V_s (**b**) compared to pure kaolinite. The V_p/V_s ratio is higher in pure smectite than in pure kaolinite (**c**)

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,).

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