

Chapter 13

Compaction of Sedimentary Rocks: Shales, Sandstones and Carbonates

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The physical properties of sedimentary rocks change continuously during burial as a response to increasing stress and temperature; they also change to a certain extent during uplift and cooling. There is an overall drive towards lower porosity with depth, which increases the sediment density and sonic/seismic velocity.

Increased effective stress from the overburden or from tectonic stress will always cause some mechanical sediment compaction (strain), expressed by the compressibility and the bulk modulus (see Chap. 11) which can be measured in the laboratory. During mechanical compaction the solids, mainly minerals, remain nearly constant so that the reduction in bulk volume is equal to the porosity loss. Sediments may also contain solid amorphous phases which are chemically unstable (e.g. opal A) The transformation of solid kerogen to fluids (petroleum) and dehydration of minerals involves however some changes in the volume of solids.

Chemically, the mineral assemblage will be driven towards higher thermodynamic stability (Lower Gibbs Free Energy) (Fig. 13.1b). These reactions involve the dissolution of minerals or mineral assemblages that are unstable, and precipitation of minerals that are thermodynamically more stable with respect to the composition of the porewater and the temperature. Higher temperatures will favour minerals with lower water content, for example by dissolving smectite and kaolinite and precipitating illite (see Chap. 4). The

rates of these reactions are controlled by the kinetic parameters such as the activation energy and thereby the temperature.

The main lithologies in sedimentary basins are shales, sandstones and carbonates, and they respond very differently to increased stress and temperature during burial.

This is important both for basin modelling and in seismic data interpretation.

There are no precise definitions for mud, mudrock and shale. The term mud is used to describe fine-grained sediment with a relatively high content of clay-sized particles, chiefly clay minerals. Carbonate mud will be discussed separately, under carbonate compaction.

The compaction (porosity loss) as a function of burial depth varies greatly because each primary lithology has a different compaction curve. While porosity may increase with depth through an interval due to changes in lithology, for each individual lithology the porosity will nearly always be reduced with depth (Fig. 13.1a).

13.1 Compaction of Mudrocks and Shales

Mudrocks and shales are often treated as one lithology in connection with basin analyses, seismic interpretation and well log analyses, but in reality they span a wide range of properties determined by the diversity of mineral composition and grain-size distribution. Furthermore, the composition of mudstones and shales changes during progressive burial due to diagenesis, which includes both mechanical and chemical compaction.

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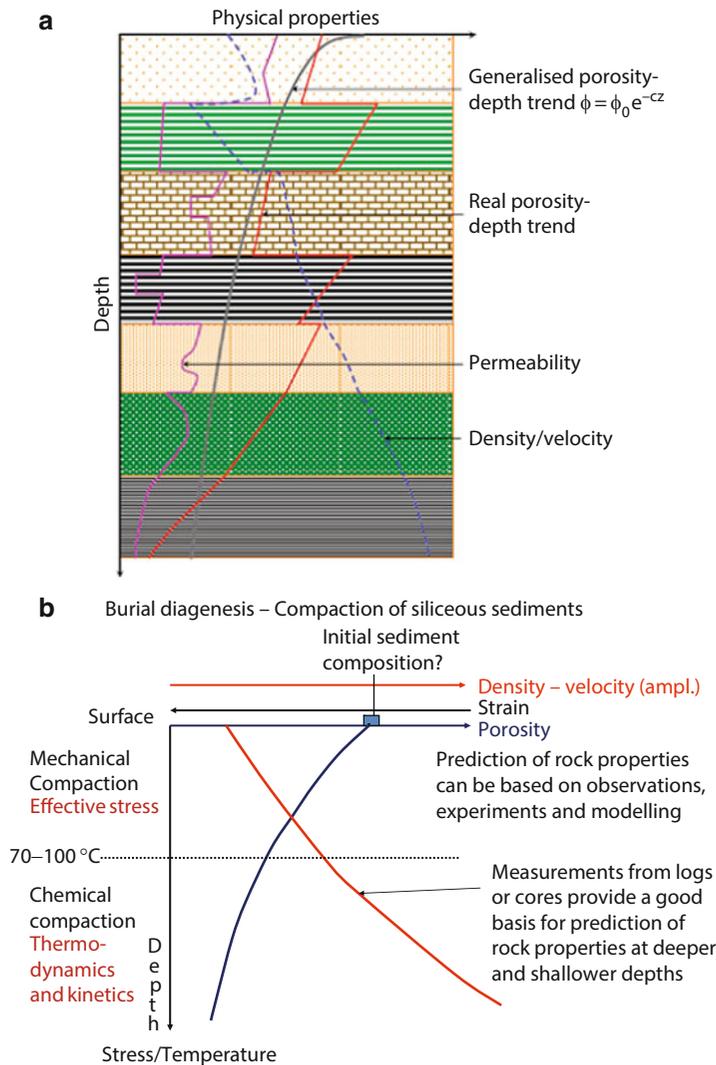


Fig. 13.1 (a) The porosity/depth trends will be different for different lithologies (primary mineralogical and textural composition). A simple exponential function may be rather far off from the real porosity/depth function. (b) Principal aspects of sediment compaction (burial diagenesis). During burial, sediments are subjected to changes in physical properties as a function of increasing stress and temperature. From an initial

sediment composition the porosity is reduced and the density and velocity are increased. Mechanically the compaction is a strain due to effective stress. Chemical compaction resulting from dissolution and precipitation of minerals is controlled in siliceous rocks by thermodynamics and kinetics and is therefore a function of temperature and time. The strain (compaction) is here independent of stress

Just after deposition the porosity of the mud near the sea or lake bottom may be extremely high, up to 70–80%. After about 1,000–2,000 m burial depth most of the mechanical compaction has taken place even if the porosity may still be relatively high (20–40%). Muddy sediments can become very compact because silt and clay can occupy much of the pore space between the larger grains, resulting in a densely packed mass.

Clay minerals, which usually account for the bulk of the finest fractions, have an impressive size range. Kaolinite particles are sheets where the longest dimension is typically 1–20 μm , while smectite particles may be smaller by a factor of 1,000 (only a few nm). Illite, chlorite and most other clay minerals have grain sizes that are intermediate between these end members. Smectite has a very high specific surface area (several hundred m^2/g) because of the small grain size and is

very sensitive to the chemical composition of the porewater. Additions of salt (NaCl or KCl) are used to stabilise soft clays for engineering purposes (construction), increasing their compressive and shear strengths. Addition of KCl in the drilling mud is also used to stabilise clays when drilling.

Sediments are often highly anisotropic and parameters like seismic velocity and resistivity can vary greatly with the orientation of the measurement relative to the bedding. Mudstones may also become increasingly anisotropic with burial depth, giving higher velocity parallel to the bedding than in the vertical direction. Experimental compaction shows, however, that the degree of grain reorientation varies markedly with the clay mineralogy and the content of sand and silt (Voltolini et al. 2009).

The composition of mudstones and shales with respect to their clay mineralogy and their content of silt and sand can provide not only important information about the environment both in and around the basin, but also about the rock properties controlling bulk compressibility, density, seismic velocity and resistivity. These parameters are important in the interpretation of seismic data and also electromagnetic surveys. This is clearly seen in some of the silty Jurassic shales from the North Sea basin (Fig. 13.2).

In the North Sea basin and Haltenbanken, poorly sorted, clayey, partly glacial sediments of Pleistocene and Pliocene age fall on a nearly linear compaction trend reaching velocities up to 2.8 km/s near the base of this sequence (Fig. 13.3). Glaciomarine clays overrun by glaciers can become very hard and compact; in the Peon gas field in the North Sea they have developed sufficiently low permeability to trap gas at just 160 m below the seafloor.

Eocene and Oligocene smectitic clays of volcanic origin have much lower velocities (<2 km/s) and densities (Fig. 13.3). Kaolinitic clay is far more compressible because it is very much coarser-grained so that the stress per grain contact is higher. Experimental compaction also shows that fine-grained kaolinite is less compressible than coarse-grained kaolinite (Fig. 11.7). Illite and chlorite are much more difficult to characterise. The clay mineral illite as determined by XRD includes both relatively coarse-grained detrital mica and very much finer-grained diagenetic illite, i.e. formed from smectite. Chlorite also varies considerably, from detrital chlorite from metamorphic rocks to authigenic, usually Fe-rich, diagenetic chlorites.

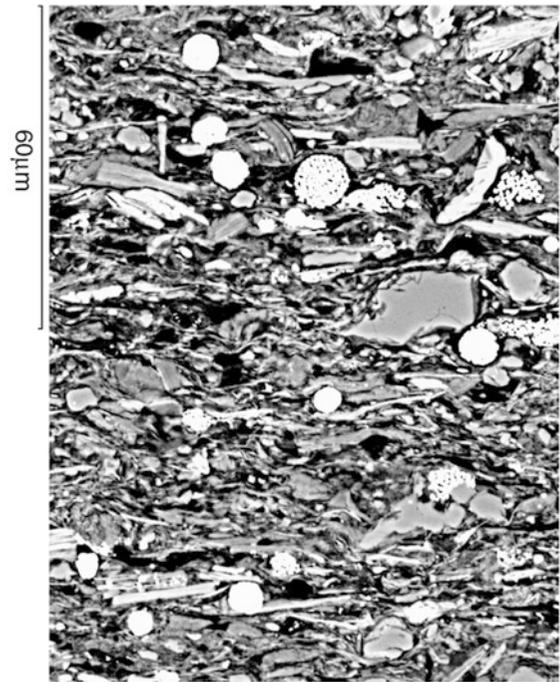


Fig. 13.2 Jurassic mudstone from the North Sea basin buried to 2.5 km depth. Note that many of the grains are of silt-sized quartz and that mica grains have a parallel orientation (scale = 0.06 mm). The white spherical structures are framboidal pyrite. The velocity in this shale is about 3 km/s (V_p 3,019– V_s 1,665 m/s)

In the laboratory the velocities (V_p and V_s) can be measured as a function of stress for mixtures of different clay minerals (see Chap. 11). Smectitic clays have very much lower velocities than kaolinitic clays but additions of silt increase the velocity. The V_s/V_p ratio also varies as a function of clay mineralogy. This is very important since this ratio is used to determine the fluid content in sand and siltstones. Mudstones and shales also may have a significant content of gas which changes the V_s/V_p ratio.

The primary composition of the mud deposited on the seafloor depends on the clay mineralogical composition and the amount of silt- and sand-sized grains. Carbonate and silica from biogenic debris are critical components with respect to burial diagenesis. Relatively moderate amounts of carbonate cement in mudstones result in high velocity at shallow depth. The source of the carbonate cement is in most cases biogenic carbonate. Fossils composed of aragonite are particularly important because they dissolve and become a source of calcite cement.

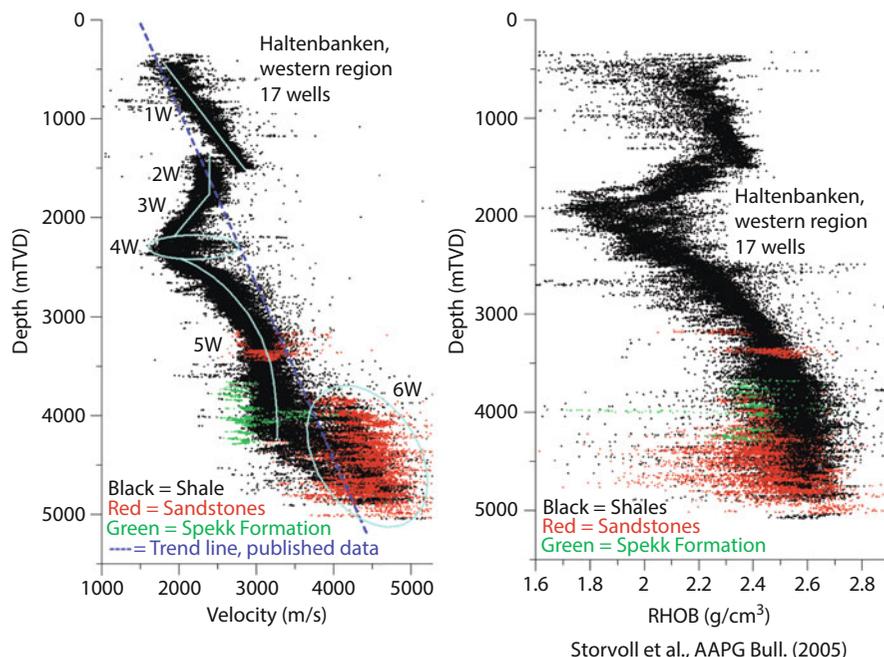
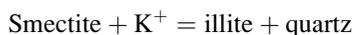


Fig. 13.3 Compaction trends are a function of burial depth and primary (initial) composition (from Storvoll et al. 2005). The poorly sorted glacially influenced Pliocene and Pleistocene sediments (1 W) compact readily while the Eocene and Oligocene smectite-rich sediments of volcanic origin (4–3 W) have low compressibility. The underlying Cretaceous and Jurassic sediments (5 W) show increases in density and velocity which probably are caused mostly by chemical compaction. Some of

the sandstones may preserve high porosity (low density) due to grain coatings and reduced quartz cementation. The Upper Jurassic Spekk Formation which is the main source (in green) is characterised by low densities and velocities (low acoustic impedance, AI). When source rocks become mature at 3–4 km depth the generation of petroleum, particularly gas, contributes strongly to this

Mudstones may contain significant amounts of organic (amorphous) silica (opal A), particularly radiolarian and diatoms, in areas with high organic productivity. Siliceous sponges may also be an important source of silica in fine-grained siltstones and sandstones.

Biogenic silica will react to form opal CT and microcrystalline quartz at about 60–80°C and this will also produce a strong stiffening of the mudstones (Thyberg et al. 2009a, Peltonen et al. 2008, Marcussen et al. 2009). Smectite becomes unstable and dissolves at temperatures above 70–100°C and mixed layer minerals and illite precipitate.



In the case of iron-rich smectite, chlorite may form as well. This reaction releases excess silica which must precipitate as quartz. Smectite is only stable

when the concentration (activity) of silica in the porewater is high. The precipitation of quartz provides a sink for the silica and lowers the silica concentration in the porewater so that the reaction can continue. Therefore the rate of quartz cementation, which is a function of temperature, controls this reaction.

It has been suggested that the silica released from the above reaction could be transported by diffusion into adjacent sandstones and precipitated as quartz cement there. Recently, small authigenic (grown in place) quartz crystals have been identified in smectite-rich mudstones which have been heated to more than about 80–85°C (Fig. 13.4). This shows that the silica is conserved locally in the mudstones. Even if the silica concentration in the mudstones were higher than in sandstones, diffusive transport in shales would be very inefficient. In mudstones without smectite or amorphous silica there are no obvious sources

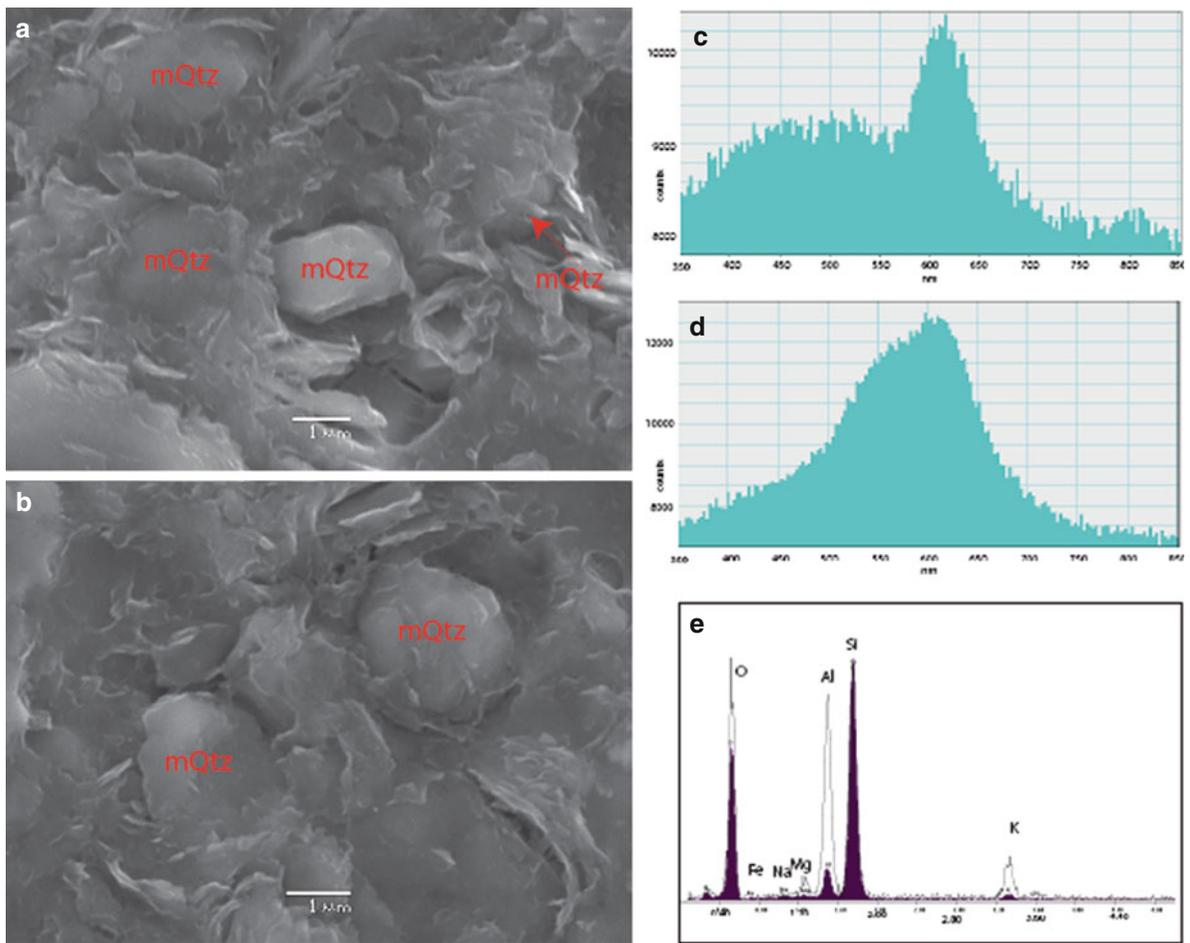


Fig. 13.4 (a and b) Authigenic microquartz precipitated in smectite-rich Late Cretaceous mudstones from the northern North Sea (from Thyberg et al. 2010). They can be distinguished from clastic quartz grains by their cathode luminescence responses (c and d) and their chemical composition (e). They

are found in mudstones which have been buried deeply enough to reach temperatures (>70–80°C) which make illite replace smectitic, providing excess silica which is then precipitated as micro-sized quartz crystals

of silica to be precipitated as early quartz cement. At greater depth most of the quartz cement is probably derived from pressure solution of detrital quartz.

In sandstones the quartz cement is sourced by solution of quartz grains at grain contacts or along stylolites, but it is not clear to what extent silt and sand grains floating in a matrix of clay will dissolve and cause precipitation of quartz as cement or as overgrowth on the grains. While quartz grains dispersed in a clay matrix may dissolve in contact with clay minerals, the surrounding clay may prevent or retard overgrowth.

At greater burial and temperatures (>130°C) kaolinite becomes unstable in the presence of K-feldspar

and releases silica which is precipitated as quartz (Bjørlykke 1981, Bjørlykke et al. 1986):



Kaolinite + K-feldspar = illite + quartz + water

This reaction is driven towards increased density (lower water content).

Kaolinite is however stable up to more than 200°C if there is no K-feldspar or other source of potassium available locally in the rock. It may then be replaced by pyrophyllite ($\text{AlSi}_2\text{O}_5(\text{OH})$) which contains less water.

Mud containing mostly quartz, illite and chlorite will be chemically stable up to high temperatures because these are metamorphic minerals. With increasing overburden and temperature, massive mudstones develop a more pronounced cleavage typical of shales.

This is due to a higher degree of parallel orientation of the sheet silicate minerals, particularly mica, illite and chlorite.

During folding, high horizontal stresses may produce an axial plane cleavage. This is controlled by a reorientation of clay minerals (sheet silicates) and also a flattening (elongation) of quartz grains by stress-driven dissolution and precipitation.

Mudstones and shales which have a high organic content contain kerogen which often occurs as thin lamina. Before the source rock becomes mature the kerogen is a part of the solid phase and can carry some of the overburden stress. When most of the kerogen is altered to oil and gas it becomes part of the fluid phase, thus changing the fluid/solid ratio so that the pore pressure reaches fracture pressure which makes expulsion more efficient. Shales with a lower organic content also will generate petroleum and gas which may migrate into the most porous and permeable layers. Some of the oil and gas, though, will be retained in the small pores as *shale gas*. There is now considerable interest in shale gas, particularly in onshore basins where drilling costs are moderate.

13.2 Sandstones

Compaction of sand and sandstones has been discussed in Chap. 4. Mechanical compaction of sand is very sensitive to primary grain size and sorting. This is a function of the depositional environment. Mechanical compaction is dominant at shallow burial down to about 2 km (70–80°C) and at greater depth compaction is mainly chemical, involving mineral dissolution and precipitation. Dissolution at grain contacts (pressure dissolution) is driven by the increased solubility due to stress causing a slight supersaturation of silica with respect to quartz and a precipitation of new authigenic quartz (overgrowth). It is now generally assumed that the precipitation, which is a function of temperature, is the rate-limiting step and that this chemical compaction is therefore rather insensitive to the stress. Dissolution at grain contacts

Sediment compaction - rock shrinkage

- Bulk modulus = Stress/strain(ΔV)
- If the strain ΔV is 0.001 or 0.1% and the bulk modulus is 50 GPa the effective stress is reduced by 50 MPa
- k volume (V_R) = Solids (V) + Fluids (porosity)
- Void ratio = $V_s/V_f = \phi/(1-\phi)$
- For isochemical reactions $V_s = \text{const.}$
- $\Delta V = \Delta\phi$, $dV/dt = d\phi/dt$

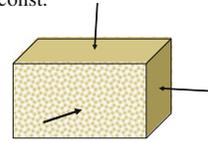


Fig. 13.5 Some definitions related to sediment compaction. During mechanical compaction the strain is produced by an increase in the effective stress. Chemical compaction in sandstones and other siliceous sediments produces strain without stress. The strain will however reduce differential stresses

occurs preferentially in contacts with mica and clay minerals which favour the development of stylolites. Transport distance between the dissolution and precipitation sites is very short and is driven by diffusion, and will be limited by the distance between the stylolites.

Compaction-driven porewater can not explain significant transport of silica. At normal geothermal gradients $3 \cdot 10^9$ volumes of water are required to precipitate one volume of quartz. In addition, porewater is generally not moving upwards in relation to the seafloor so there is little cooling of the porewater (see Chap. 4).

As in mudstones, chemical compaction in sandstones is mostly controlled by temperature and both sandstones and mudstones compact chemically during burial. Overpressure reduces the effective stress but has then little effect on compaction.

The loss of porosity results in higher density and a reduction in the total rock volume or shrinkage (Fig. 13.5). Even a very small loss of porosity (strain) by chemical compaction will reduce the bulk volume so that the stress is reduced.

This shrinkage will contribute to a reduction in horizontal stress because some of the compaction may occur in the horizontal direction. This is indicated by the leak-off pressures at greater depth (Fig. 13.6). This may reduce horizontal tectonic stresses.

In the upper parts of sedimentary basins (<70–80°C) the compaction of siliceous sediments follows the laws of soil and rock mechanics. At greater

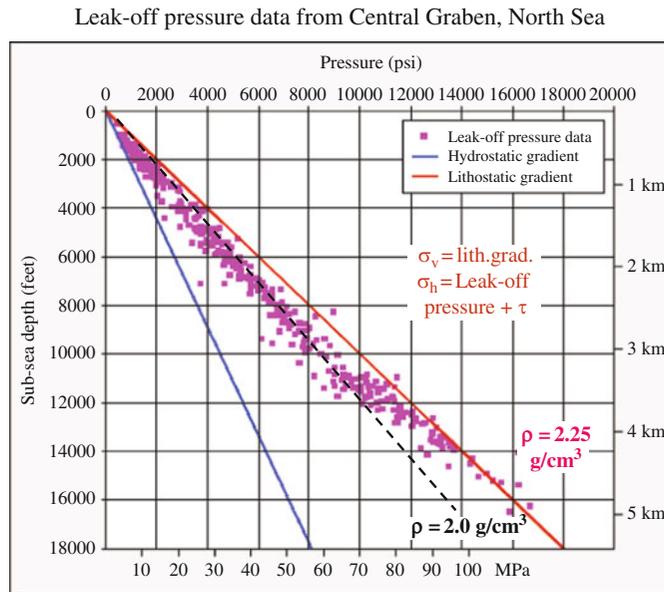


Fig. 13.6 Leak-off pressure is an indication of the horizontal stress and at 3–4 km this is nearly equal to the vertical stress. This suggests that during chemical compaction the rocks compact both *vertically* and *horizontally*, thus reducing differential

stress. At very slow strain rates a sandstone may respond nearly as a fluid where the stress is equal in all directions. The horizontal stress is close to the vertical overburden stress. From *The Millennium Atlas*. Geological Society of London, 2003

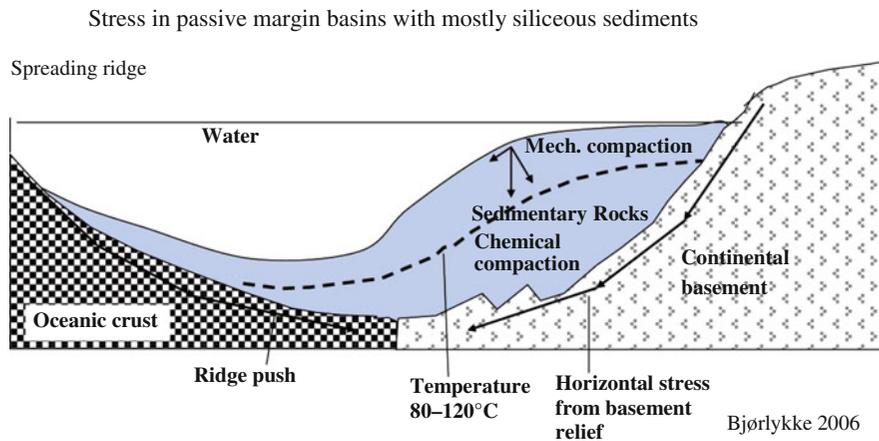


Fig. 13.7 Simplified cross-section through a sedimentary basin on a passive margin. Most of the tectonic stress is transmitted through the basement and the well-cemented sedimentary rocks.

In the case of ice loading, the strain rates are relatively high and the response in the sediments will be mostly mechanical compaction. Gravitational stress may also be important

depth compaction is mainly chemical and controlled by temperature (Fig. 13.7).

When the compaction is mechanical any reduction in the effective stress due to uplift or the build-up of overpressure will cause the sedimentary rocks to become overconsolidated, and the deformation does not follow the virgin loading curve (see Chap. 11).

Chemical compaction due to quartz cementation in siliceous sediments will continue during uplift as long as the temperature is higher than 70–80°C. Even a small shrinkage (porosity loss) due to this compaction will result in a reduction in the *in situ* stress controlled by the bulk modulus. The mechanical extension due to unloading will then at least partly be compensated for

by chemical compaction, and open fractures will gradually be healed by quartz cement.

13.3 Carbonate Compaction

Compaction of carbonates is controlled by principally very different processes than in siliceous sediments. Because the kinetics of carbonate dissolution and precipitation are so much faster than for siliceous rocks (mudstones and shales), temperature is not the main control on carbonate compaction. Cementation of carbonate sediments into hard solid rocks may occur right near the surface. In addition the presence of aragonite which is thermodynamically less stable than calcite provides a strong drive for cementation. The sediments then become mechanically overconsolidated and may be unable to undergo further mechanical compaction even when subjected to 40–50 MPa (4–5 km depth). Carbonate sediments like the Chalk, composed almost entirely of low-Mg calcite, undergo little dissolution and cementation also along stylolites at 2–3 km depth. Overpressure is very important in reducing both mechanical compaction and pressure dissolution. In the Ekofisk Field, Chalk

may have porosities exceeding 30% at nearly 3 km burial depth due to high overpressure, because the effective stress only corresponds to about 1 km without overpressure.

The processes controlling porosity loss in carbonate sediments are still poorly understood.

The dissolution rate may be more important compared to sandstones. At the contact between two calcite grains there is probably only a very thin layer of water, while clay minerals have a double layer due to the negative surface charges. The transport of calcium along the calcite grain contacts may also be rate-limiting. Carbonate grains, particularly of fossils, may have an organic coating which may influence precipitation.

Early porosity reduction in carbonate at shallow depth may help to preserve the resultant porosity during deeper burial. Carbonates have generally lower porosity than sandstones at the same depth (Fig. 13.8) but there is a wide range of porosity/depth values, particularly for carbonates.

Near the surface where there may be meteoric water flow the system is relatively open and net porosity may be created by dissolution. There is, however, limited potential for mass transport of carbonate in

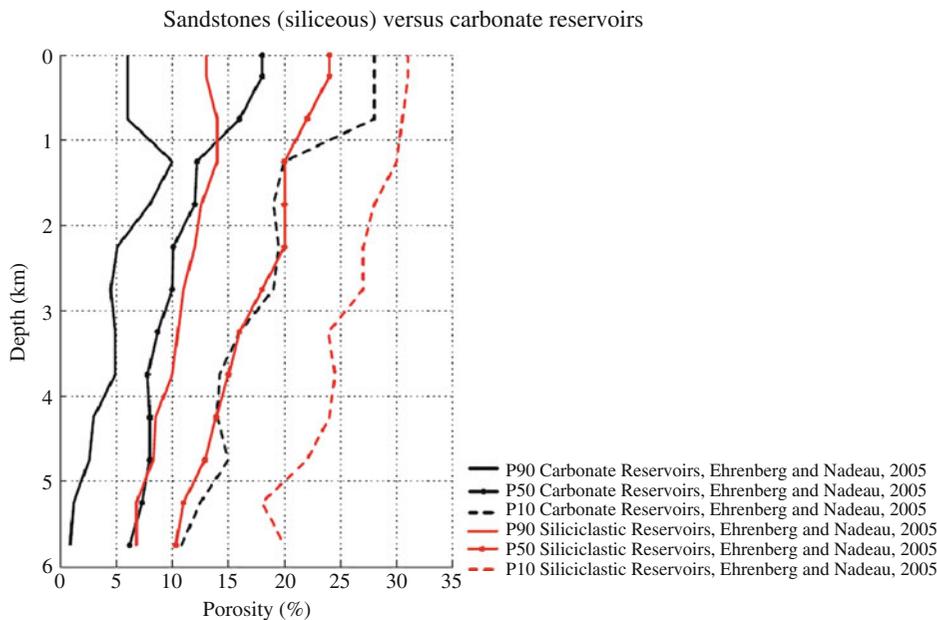


Fig. 13.8 Compaction trends for carbonates and sandstones (from Ehrenberg and Nadeau 2005). Average porosity versus top depth for global petroleum reservoirs. P90, P50, and P10

indicate that 90, 50 and 10% of the reservoirs' values have higher porosity than this value

solution during burial and the reactions must be nearly isochemical. This is because the porewater will always be closely in equilibrium with the carbonate minerals that are present. This leaves little potential for transport by diffusion, or by advection, because the flow rates are so small, particularly in relation to the isotherms. Focused flow as along faults will cause some dissolution because of the retrograde solubility of carbonates like calcite. The solubility is also a function of pressure but flow across pressure barriers is rather limited.

Much of the compaction of carbonate rocks occurs along stylolites because the dissolution and transport along grain contacts are enhanced by the presence of sheet silicates (see [Chap. 5](#)).

13.4 Summary

Shales, sandstones and carbonates follow different compaction trends and they are controlled by principally different processes.

Both shales and sandstones compact mechanically as a function of effective stress until chemical compaction takes over and further compaction is mainly a function of temperature and time. The initial mineralogical and textural composition is very important both for sandstones and mudstones (shales). This is clearly shown by experimental compaction of sediments with different mineralogical and textural composition. Smectitic clays are very different from clays with other types of clay minerals.

Carbonate sediments may compact chemically at very shallow depth and low temperature and the compaction process is driven by a complex interaction between stress and chemical compaction, but the temperature is less important than in siliceous sediments. One of the main factors controlling compaction and rock properties in carbonates is the primary content and distribution of aragonite, causing early cementation.

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