Chapter 4

Sandstones and Sandstone Reservoirs

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4.1 Introduction

About 50% of all petroleum reservoirs are sandstones; outside the Middle East, carbonate reservoirs are less common and the percentage is even higher. The most important reservoir properties are porosity and permeability, but pore geometry and wetting properties of the mineral surfaces may also influence petroleum production. Sandstones provide reservoirs for oil and gas, but also for groundwater which is a fluid that is becoming increasingly valuable. Sandstones are deposited in many different sedimentary environments by marine, fluvial and eolian (wind) processes.

The outer geometry and distribution of sand bodies is determined by the depositional environment and the reservoir properties. The internal properties (porosity, permeability) are, however, critical for petroleum recovery.

The properties of sandstone reservoirs are functions of the primary composition, which is controlled by the textural and mineralogical composition (provenance) of the depositional environment and by the diagenetic processes near the surface and during burial.

Sand and sandstones are rocks which consist largely of sand grains, i.e. sedimentary particles between 1/16 and 2 mm in diameter. However, sandstones also contain greater or lesser amounts of other grain sizes and we find transitions to more silt- and clay-rich rocks. Most sandstones have a well-defined upper grain-size limit with variable contents of silt and clay. If they have a significant content of coarser grains (pebbles) we call them conglomeratic sandstones. Most classification systems are based on the relationship between the relative quantity of sand-sized grains, the composition of the sand grains, and the clay and silt content (matrix).

If we use a four-component diagram we can distinguish between clay, and sand grains which consist of quartz, feldspar and rock fragments (or unstable rock fragments, U.R.F.) (Fig. 4.1). Rock fragments consisting of microcrystalline (or cryptocrystalline) quartz (including chert) are usually classified together with the quartz mineral grains. Sandstones with more than 25% feldspar and a low content of rock fragments are called *arkoses*. If the percentage of rock fragments is high, we speak of *lithic sandstones* which are normally derived from very fine-grained sedimentary rocks or basalts and intrusive igneous rocks where one sand grain often consists of several minerals.

Quartz arenite or *orthoquartzite* are the terms for pure quartz sandstones which contain less than 5% feldspar or rock fragments. Sandstones with moderate feldspar contents (5–25%) are called *subarkoses*. When granitic rocks and other coarse- to medium-grained rocks are broken down by weathering or erosion, they form sand grains consisting for the most part of single minerals, mostly quartz and feldspar.

The prerequisite for forming arkose is that not too much of the feldspar in the source rock has been weathered to clay minerals (e.g. kaolinite). Arkoses are therefore formed if granites and gneisses are eroded rapidly in relation to weathering, and the sediments are buried in a basin after a relatively short sediment transport. In the great majority of cases arkose is therefore associated with sedimentary rift basins formed by faulting in gneissic and granitic

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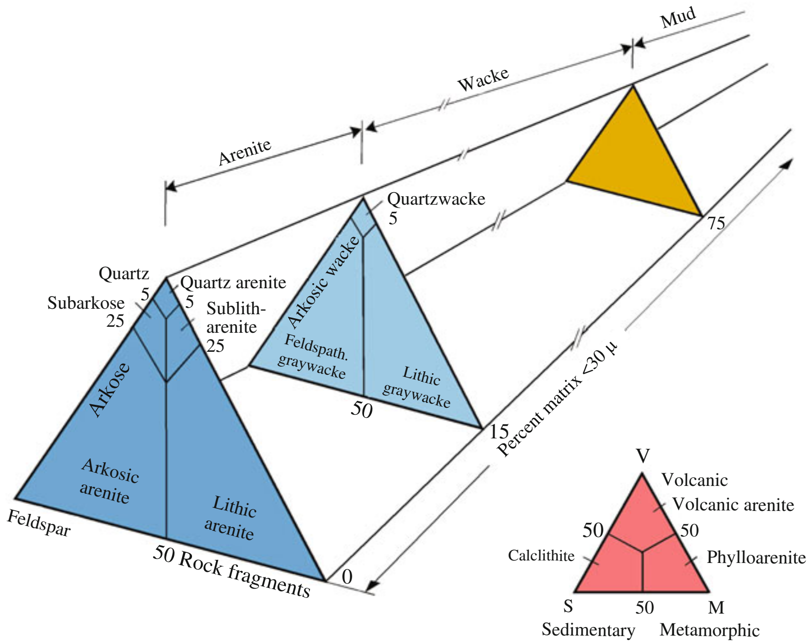

Fig. 4.1 Classification of sandstones (from Dott 1964)

rocks, i.e. continental crust. In cold climates the rate of weathering is reduced, thus preserving feldspar and unstable rock fragments.

In tectonically stable areas with mature relief there is far more time for weathering of both *in situ* bedrock and sediment in transit. In such environments sand particles (grains) are deposited and eroded many times before reaching their final deposition site. A greater proportion of the feldspar grains will then break down during transport and subarkoses or orthoquartzites will be deposited. Orthoquartzites (or quartz arenites) are the purest quartz sandstones, formed when weathering has eliminated virtually all the unstable minerals to leave a concentration of quartz and some heavy minerals. This is particularly true of beach sand where nearly all the clay particles have been removed. Sand deposited by rivers (fluvial sand) is less well sorted and may contain more clay. Sandstones with more than 15% matrix are called *greywackes*.

Sand which is transported in suspension or through mass flow (e.g. turbidites) can have poorer sorting, a high matrix content, and grade into sandy mudstones and form greywackes. Clay minerals in sandstones may form after deposition during diagenesis from alteration of feldspar, mica and rock fragments. This makes the sediments less well sorted than at the time of deposition.

Basic rocks like gabbro and basalt and minerals like amphibole and pyroxene are inherently unstable both mechanically and chemically. Primary sand grains of volcanic or basic rocks may break down to become part of the matrix and it is then difficult to distinguish this material from the primary matrix. Greywacke is therefore typical of areas where the sand grains are derived from volcanic or basic rocks along converging plate boundaries (fore-arc, inter-arc, back-arc basins). Weathering of basic rocks will produce nearly exclusively clay since there are no quartz grains.

We have seen that the various types of sandstone reflect different source rocks and areas with varying tectonic stability. Studies of different types of sandstone and their mineralogical maturity can therefore be used as palaeo-indicators of relief and climate, and also of tectonic deformation in the geological past.

4.2 Prediction of Reservoir Quality

The properties of all reservoir rocks are continuously changing, from the time the sediments are deposited through to their burial at great depth and during any subsequent uplift. This is a combined function of mechanical compaction and of chemical processes involving dissolution and precipitation of minerals.

At any given burial depth the properties depend on the composition of the sandstones when at shallow depth, and on their temperature and stress history during burial. Practical prediction of the porosity and permeability during exploration and production is only possible if the processes that change these parameters are understood.

It should be realised that the starting point for the diagenetic processes is the initial sandstone composition. This is a function of the rocks eroded (provenance), transport, and depositional environments. Diagenetic models must therefore be linked to weathering and climate, sediment transport, facies models and sequence stratigraphy, and should be integrated in an interdisciplinary *basin analysis*.

Diagenesis is often considered a rather specialised field of sedimentology and petroleum geology, but it embraces all the processes that change the composition of sediments after deposition and prior to metamorphism. The most important factor in predicting reservoir quality at depth is the primary clastic composition and the depositional environment (Fig. 4.2). The diagenetic changes also determine the physical properties of sandstones, such as seismic velocities (V_p and V_s) and the compressibility (bulk modulus, see Chap. 11). This is also critical when predicting physical rock changes during production (see 4D seismic, Chap. 19).

The main diagenetic processes are:

(1) Near-surface diagenesis. Reactions with fresh groundwater (subsurface weathering). In dry environments, with saline water concentrated by evaporation. Sand may also be cemented with carbonate cement near the seafloor.
(2) Mechanical compaction, which reduces the porosity by packing the grains closer together and by grain deformation and fracturing, increasing their mechanical stability. Mechanical compaction is a response to increased effective stresses during burial and follows the laws of soil mechanics.
(3) Chemical diagenesis (compaction), which includes dissolution of minerals or amorphous material and precipitation of mineral cement so that the porosity and the rock volume are reduced. The clastic minerals in the primary mineral assemblage are not in equilibrium, and there is always a drive towards thermodynamically more stable mineral assemblages. Kinetics determine the reaction rates, which for silicate reactions are extremely slow so temperature plays an important role.

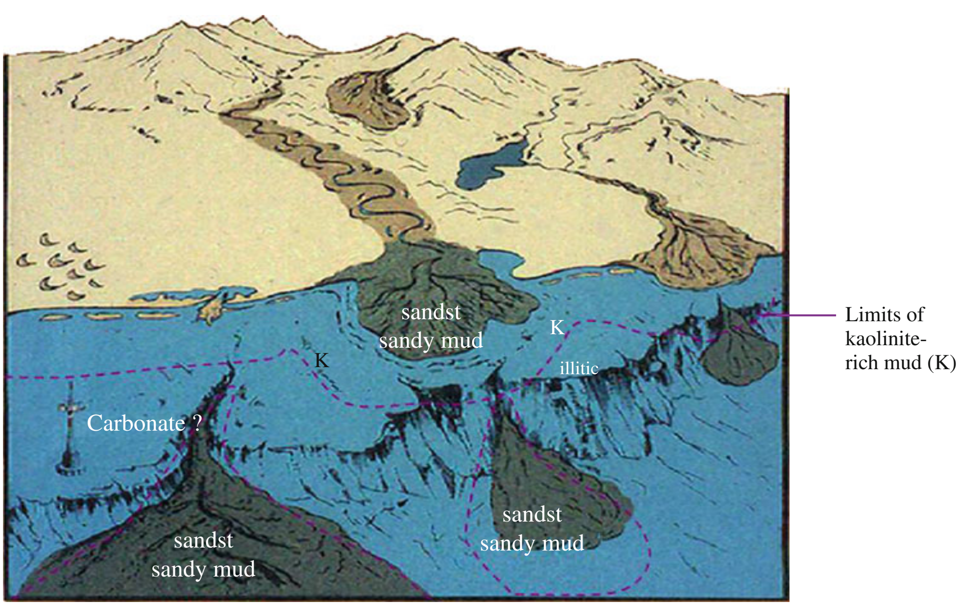

Fig. 4.2 Schematic illustration of a sedimentary basin on a continental margin. The primary composition of the sediments is a function of the provenance, transport and depositional environment. Fluvial, deltaic and shallow marine sediments will be flushed by meteoric water after deposition, particularly in humid climates

(4) Precipitation of cement (e.g. quartz cement) will increase the strength of the grain framework and prevent further mechanical compaction. The sandstone is then overconsolidated – not due to previously higher stress, but due to cementation. This is sometimes referred to as pseudo-overconsolidation. Further compaction will then mostly be controlled by the rate of dissolution and precipitation.

In the following the diagenetic processes typical of different burial depths will be discussed.

4.3 Early Diagenesis

As soon as sediments are deposited, early diagenetic reactions start to modify the primary sediment composition. At very shallow burial depth (<1–10 m), sediments have the maximum potential to react with the atmosphere or water, both by fluid flow and diffusion. Transport of dissolved solids by diffusion and fluid flow (advection) is most efficient near the surface; in the case of diffusion within about 1 m of the seabed. The potential for sediments to change their bulk composition after deposition is therefore much higher at shallow depth than at greater burial. Near the surface on land, and also in the uppermost few centimetres of the seabed, the conditions may be oxidising, while at greater depth in the basin they are always reducing.

Precipitation of minerals due to porewater concentration by evaporation can only occur on land or at shallow depth within enclosed basins (see Chap. 6).

On land, sediments are exposed to air and fresh (meteoric) water. Weathering is partly due to reactions with oxygen in the atmosphere and partly due to dissolution of minerals in freshwater, which is initially undersaturated with respect to all the minerals present. These are soil-forming processes which can be considered to be examples of early diagenesis.

In desert environments groundwater and occasional rainwater may become concentrated through evaporation, causing precipitation of carbonates and also silicates. Coatings of red or yellow iron oxides and clays frequently form on desert sand and this may subsequently retard or prevent quartz cementation at greater depth.

In the sea, the water above the seabed is normally oxidising. Only where there is poor water circulation (poor ventilation) is the lower part of the water column likely to be reducing, though the phenomenom is more widespread in lakes and inland seas like the Black Sea. However, even below well-oxygenated water, oxidising conditions extend in most cases for only a few centimetres into the sediments, since oxygen is quickly consumed by the oxidation (decay) of organic matter in the sediment. This is for the most part facilitated biologically by bacteria. Accumulating sediments normally contain sufficient organic matter to serve as reducing agents in the porewater. This organic matter is comprised of both the remains of bottom fauna and of pelagic organisms, including algae, accumulating on the seafloor, and also often includes terrestrial plant debris transported into the basin.

4.4 Redox-Driven Processes on the Seafloor

Across the *redox boundary* there is a high gradient in the concentration of oxygen and sulphate, and of ions that have very different solubilities in oxygenated and reduced water. The redox boundary is usually just 1–20 cm below the seafloor and represents equilibrium between the supply of oxygen by diffusion, and its consumption by the (mostly biological) oxidation of organic matter. The oxygen content in the porewater thus decreases rapidly below the water/sediment interface, providing a concentration gradient for the downward diffusion of oxygen into the uppermost sediments.

The rate of downward diffusion of oxygen is a function of the concentration gradient of oxygen in the porewater and the diffusion coefficient in the sediments. The diffusion coefficient in coarse-grained sand is higher than in mud and therefore sand tends to have a deeper redox boundary than mud.

Oxygen can also be consumed in the sediments by the oxidation of elements like iron and manganese in minerals, but this is rare in marine environments and more common in continental deposits. In most marine environments there is enough organic matter to serve as reducing agents and therefore little oxidation of iron in minerals takes place, which explains why marine sediments do not normally acquire a red colour. A notable exception is red oxidised mud which may form in marine environments characterised by slow sedimentation rates and low organic productivity. These muds are not very common but occur in some deep-water facies and also in shallower water environments with low sedimentation rates.

Uranium is highly soluble in seawater as uranyl (UO^{2+}) and precipitates as reduced uranium oxide (UO_2) on organic matter in the water column and below the redox boundary. There is thus a strong concentration gradient transporting uranium from the seawater down into the sediments. The adsorption of uranium onto organic matter settling on the seafloor coupled with restricted ventilation of the water above the sediment at the time of deposition, makes source rocks like the Kimmeridge shale strongly enriched in uranium, causing peaks on the gamma ray well log curve.

Iron and manganese may be transported upwards through the sediments in the reduced state by diffusion and then precipitate on the seafloor as oxides because of their reduced solubility in the oxidised state. Iron may also be precipitated below the redox boundary as iron sulphides or iron carbonate (siderite), though iron carbonates will not form during sulphate reduction. This is because all the iron will be precipitated as sulphide, which has a much lower solubility than siderite. Manganese is not precipitated as sulphides because of the high solubility of Mn-sulphides, but may be precipitated as Mn-carbonate in the reduced zone.

At a depth where there is practically no more free dissolved oxygen in the porewater, sulphates can be used by sulphate-reducing bacteria. The reduction of sulphates produces sulphides such as pyrite.

The composition of clastic sediments is modified by the addition of new components produced locally within the basin:

(1) Biogenic carbonates and silica.
(2) Authigenic minerals precipitating near the seabed such as carbonates, phosphates, glauconite, chamosite, sulphides and iron and manganese minerals.
(3) Meteoric water-flushing causing leaching of feldspar and mica and precipitation of kaolinite beneath the seafloor.

4.5 Importance of Biogenic Activity

Bioturbation plays an important role in changing the textural composition of the sediments after deposition. The burrowing organisms eat mud and thereby oxidise organic matter and physically destroy the primary lamination. Sediments overturned by bioturbation become more exposed to oxidation at the sea or lake bottom. Bioturbation may reduce the porosity and permeability of sandy laminae by mixing clay with clean sand.

Bioturbation will also destroy thin clay laminae, which may significantly increase the vertical permeability and this may be very significant for reservoir quality. Undisturbed primary lamination may be evidence of rather rapid sedimentation giving little time for a burrowing bottom fauna to become established, or alternatively indicate strongly reducing conditions restricting the fauna. Black shales usually have well preserved lamination due to lack of burrowing organisms. The presence or absence of burrowing also influences the physical properties, particularly the difference in velocity and resistivity parallel and vertical to bedding (anisotropy) and this may be important for geophysical modelling.

Burrowing worms produce faecal material which may develop into smectite-rich clays, which in turn may develop into chlorite coatings, thus improving reservoir quality. Early diagenetic formation of coatings on quartz grains is extremely important due to its role in preserving porosity at greater depth.

Most clastic depositional environments have some organisms producing organic matter which, at least in part, is incorporated within the sediments. Both sandstones and mudstones nearly always contain significant amounts of biogenic material from calcareous, and sometimes also siliceous, organisms and this may later be an important source of carbonate and silica cement at greater burial depth.

Marine organisms composed of aragonite dissolve during relatively shallow burial and calcite precipitates either as replacements within the fossils (neomorphism) or as cement in pore space between the grains.

Carbonate cement in sandstones may form layers or concretions and in most cases is derived from biogenic carbonate, particularly from organisms composed of aragonite. Siliceous organisms composed of opal (e.g. diatoms or siliceous sponges) may be an important source of microquartz coatings on quartz grains at greater depth.

Carbonate cements in both mudstones and sandstones are mostly due to dissolution and reprecipitation of biogenic carbonate or early aragonite cement. There are usually no other major sources of carbonate cement. In the sulphate-reducing zone, carbonate concretions form, often with a negative $\delta^{13}C$ due to the CO_2 produced during sulphate reduction. Carbonate concretions in cores may be mistaken for continuous carbonate layers but it is possible to recognise that they are concretions (Walderhaug and Bjørkum 1998). Even if CO_2 is generated from organic matter, there are few Ca^{2+} sources available in sandstones or mudstones for making calcite. Leaching of plagioclase can supply some Ca^{2+}, which can be precipitated as calcite, but this can only account for very small amounts of the calcite observed in such sediments. The distribution of carbonate cement is related to facies and sequence stratigraphy.

The evolution of pelagic planktonic calcareous organisms in the Mesozoic drastically increased the supply of carbonate on the seafloor, including in deeper waters. Before then most of the carbonate was produced by benthic organisms restricted to shallow water facies. Upper Jurassic and younger sandstones often contain abundant calcite cement due to the 'rain' of calcareous algae, foraminifera and other planktonic organisms settling on the seafloor. Silica-producing organisms may also be important for diagenesis and reservoir quality at greater burial. Organisms like siliceous sponges are composed of amorphous silica which at higher temperatures will be dissolved and replaced by opal CT and quartz. Diatoms and radiolarians may also be a major source of silica which will be precipitated as quartz. Diatoms appeared during the Cretaceous and have been a major source of amorphous silica during the Cenozoic. Diatoms can produce pure siliceous rocks like the Tertiary Monterey Fm of California, which is both a source rock and a fractured reservoir rock.

Biogenic carbonate is in most cases the main source of calcite cement. The distribution of such cement must therefore be linked to sedimentary facies, more specifically to biological productivity relative to the clastic sedimentation rate. Environments with low clastic sedimentation rates, particularly submarine highs, often have high organic carbonate production.

4.6 Meteoric Water Flow and Mineral Dissolution

Meteoric water is rainwater which infiltrates the ground. Initially this water is distilled water and therefore undersaturated with respect to all minerals. The reactions between meteoric water and the land surface are an important part of the weathering process. Rainwater contains carbon dioxide (CO_2) and sulphur dioxide (SO_2) from the air and is therefore slightly acidic, producing carbonic acid (H_2CO_3) and sulphuric acid (H_2SO_4).

Some of the rainwater seeps down to the groundwater, and as long as the groundwater table is above sea level, meteoric water will flow along the most permeable beds into the basin. Meteoric water will first dissolve carbonates and then slowly dissolve unstable minerals like feldspar and mica (Fig. 4.3).

Decaying organic matter in the ground produces CO_2 which is added to the groundwater, making it more acid. Humic acids generated by decaying plants also hasten the weathering reactions. At the same time this acidity is neutralised by weathering reactions with silicate minerals like feldspar and the dissolution of carbonates which consume protons (H^+). As the

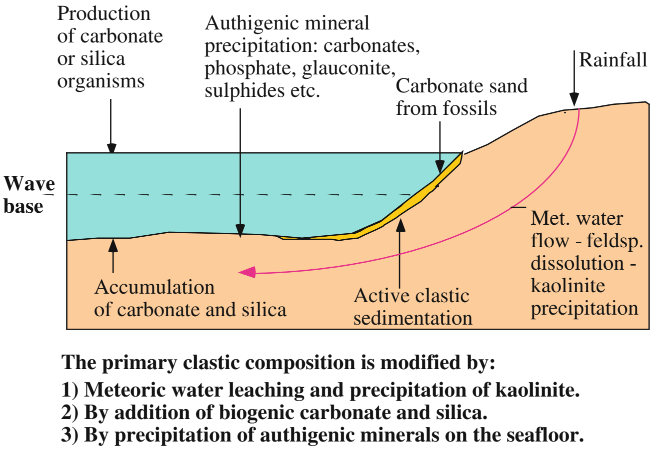

Fig. 4.3 Diagenetic processes in shallow marine environments. Sandstones deposited in these environments will be flushed by meteoric water flow and/or from the delta top, causing dissolution of feldspar and mica. Calcareous fossils and early carbonate cement may be a very important addition to the composition of the sandstones. The occurrence of siliceous organisms such a sponges can strongly influence reservoir quality at depth

groundwater reacts with minerals and in some cases with amorphous phases, it will approach equilibrium with many of the minerals present and this will happen first with carbonates. In the case of silicate minerals these reactions are very slow so the porewater may remain under- or supersaturated for a long time with respect to silicate minerals like quartz and feldspar.

Depending on the elevation of the groundwater table and the distribution of permeable layers (sandstones), the flow of meteoric water can extend beneath the seafloor far out into sedimentary basins. Reactions between meteoric water and minerals occur in the ground and are a kind of subsurface weathering along the groundwater flow paths. Leaching by meteoric water is generally strong in fluvial and alluvial sediments. Even within dry river beds there is a focused flow of groundwater.

The groundwater level represents the head (potentiometric surface) for groundwater flow and groundwater therefore has a potential to flow through sediments or other aquifers far below sea level.

The rates of leaching of minerals like feldspar and mica and the precipitation of kaolinite are functions of the flux of groundwater flowing through each rock volume per unit of time. These are in principle weathering reactions similar to those which take place during normal weathering in a humid climate. Cations like Na^+ and K^+ are stripped from silicate minerals like feldspar and mica and brought into solution.

feldspar often remains and the dissolved aluminium and some of the silica is precipitated as kaolinite, so there is a rather small increase in porosity and reduced permeability (Fig. 4.4a). The pores between the kaolinite crystals may be too small (Fig. 4.4b, c) to be filled with oil so that the oil saturation is reduced in kaolinite-rich sandstones (see Chap. 20).

The silica released from feldspar dissolution can normally not be precipitated as quartz because of the low temperature near the surface, but remains in solution even if the porewater is highly supersaturated with respect to quartz. Silica must, however, also be removed along with potassium by the flowing water. If the silica concentration in the porewater increases too much, kaolinite is no longer stable and smectite will precipitate instead. This happens in sediments rich in volcanic material or biogenic silica and where the flux of meteoric water is low. In a desert environment evaporation of groundwater may increase the silica concentration and make smectite more stable.

The porewater does not have to be acidic for kaolinite to form, but the K^+/H^+ ratio must be low. If the pH is high the K^+ concentration has to be correspondingly lower. Authigenic kaolinite may also form in impure limestones as a result of meteoric water flushing, and the porewater is then certainly not acidic. Even if there is only a small amount of carbonate it will buffer the composition of the porewater.

The average groundwater flux is determined by the rainfall and the percentage of water infiltration into the

These reactions can be written as below:

$$2K(Na)AlSi_3O_8 + 2H^+ + 9H_2O = Al_2Si_2O_5(OH)_4 + 4H_4SiO_4 + 2K^+(2Na^+)$$

Feldspar — Kaolinite — Dissolved silica and cations

$$2KAl_3Si_3O_{10}(OH)_2 + 2H^+ + 3H_2O = 3Al_2Si_2O_5(OH)_4 + 2K^+$$

Muscovite — Kaolinite

We see from these reactions that low K^+/H^+ ratios will drive the reactions to the right. Dissolution of feldspar and mica and precipitation of kaolinite require that the reaction products, Na^+, K^+ and silica, are constantly removed and that there is a supply of new freshwater which is undersaturated with respect to feldspar and mica. Without a through flow of water these reactions stop because the reaction products on the right hand side of the equations are not removed. Groundwater must flow into the ground and up to the surface again or on to the seafloor. A clay coating on

ground. In moderately humid climates the rainfall may be 1 metre/year and the infiltration in the order of 0.1–0.3 metres/year. High-permeability subsurface pathways (aquifers) focus the flow. The aquifers may be sand or gravel beds in muddy sediments. Meteoric water may penetrate deeply into sedimentary basins because of the potential created by the head of the groundwater table but the flux of meteoric water decreases strongly in the deeper parts of basins.

The meteoric water will gradually become less undersaturated with respect to minerals like feldspar

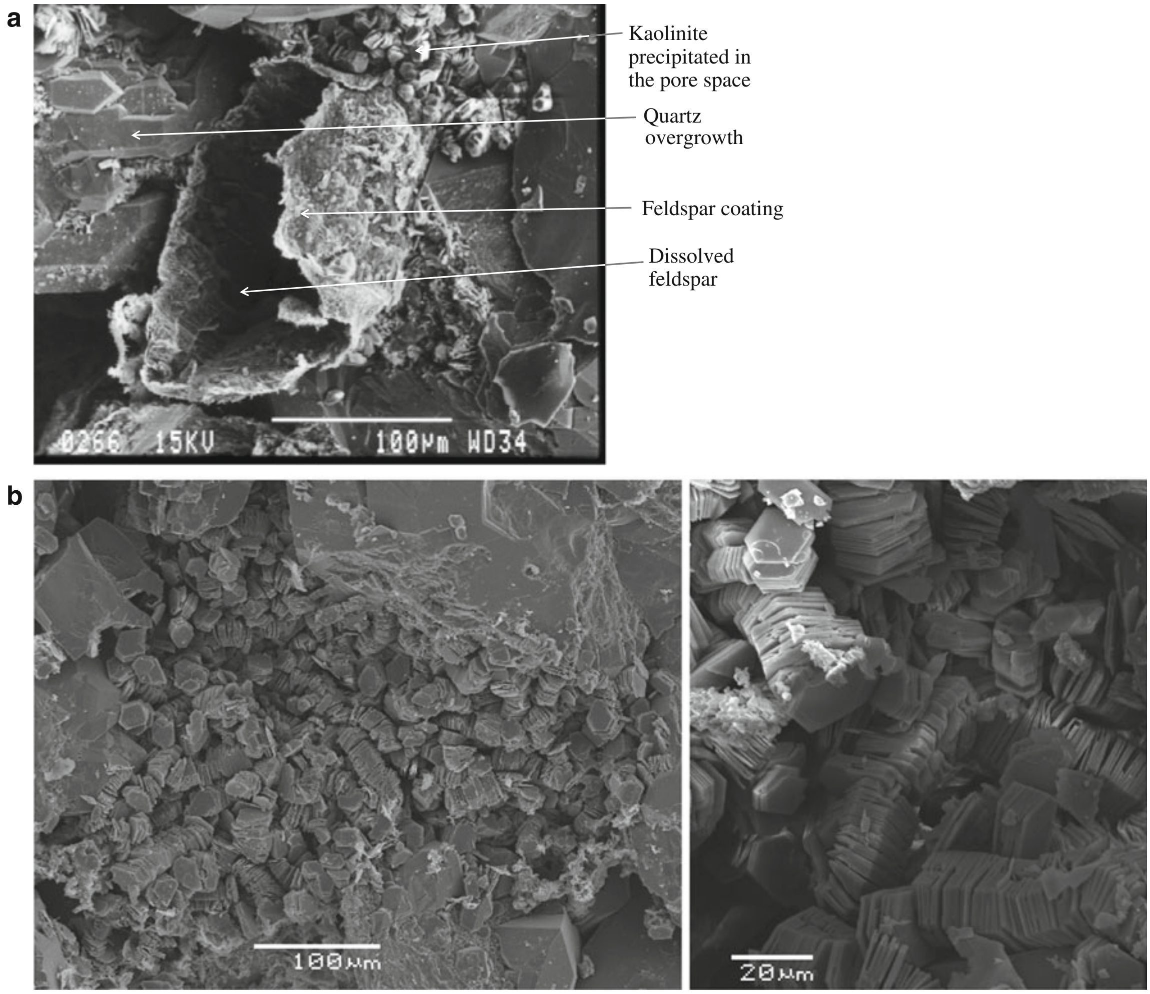

Fig. 4.4 (**a**) Scanning electron microscope picture of a sandstone (Brent Group) from the North Sea. The scale is 0.1 mm (100 µm). In the *centre* of the picture we see a cavity left by a dissolved feldspar grain. A clay rim around the feldspar remains undissolved, outlining the primary grain morphology. In the *upper part*, authigenic kaolinite crystals are forming small (10–20 µm) booklets. They have formed from the silica and aluminium released when the feldspar was dissolved by meteoric water. To the *left*, authigenic quartz is growing on clastic quartz. Note the relatively large pores between quartz and feldspar grains and the small pores between kaolinite crystals. (**b**) Pore-filling authigenic kaolinite. We see that the pores between the kaolinite crystals are very small – only 1–2 µm (from T.E. Maast unpublished)

and mica and its leaching capacity will gradually diminish. In the North Sea basin Middle Jurassic sandstone has been uplifted and eroded during the late Jurassic, and later overlain by Cretaceous sediments. Evidence of meteoric water leaching is however limited to a few metres below the unconformity (Bjørkum et al. 1990). The most intense mineral leaching will therefore occur near the surface or at relatively shallow depth beneath the seafloor. In areas with low sedimentation rates the total flow of water through the sediment will be higher because the sediments remain longer at shallow depth. If the sediment stays in the zone intensively flushed by meteoric water, the amount of feldspar leaching will be high. In basins with high sedimentation rates, synsedimentary faulting (i.e. growth faults) may disconnect sand bodies from the main freshwater aquifers. The degree of feldspar leaching could then be used as an indication of the conductivity in the reservoir.

River water and groundwater are usually supersaturated with respect to quartz but undersaturated with respect to amorphous silica. About 10–30 ppm dissolved silica is common in groundwater and shallow porewater while the solubility of quartz at 20°C is only 4–5 ppm, showing that quartz does not form at low temperatures. In very alkaline water (i.e. E. African lakes), quartz may precipitate at low temperatures.

The early burial history of sandstones is not well studied for the simple reason that cores are not normally taken at depths shallower than 1–1.5 km in offshore basins, while onshore, erosion may have removed most of the youngest sequence. Looking at sandstone thin sections one can often get the impression that kaolinite is precipitated at a relatively late stage because it is a pore-filling mineral that may subsequently be surrounded by quartz cement. More detailed textural studies and isotopic evidence indicate that the kaolinite is formed early and may be enclosed in quart cement. Pore-filling kaolinite must, however, have been hanging to the pore wall and may have been pushed aside by growing quartz cement.

The isotopic composition ($\delta^{18}O$) of kaolinite suggests that it precipitated at relatively low temperatures, in the range 30–60°C, depending on the assumptions made about the isotopic composition of the porewater (Glasmann 1989). These temperatures are little higher than should be expected during meteoric water flushing and it is possible that some of the kaolinite may be recrystallised at a higher temperature, resetting the isotopic composition. Much of what has been described or analysed as kaolinite has turned out to be dickite, which has the same composition but often with thicker, more blocky crystals. Studies have shown that dickite often replaces some of the kaolinite when temperatures exceed 100°C.

Another possibility is that kaolinite may form diagenetically from other precursor minerals such as gibbsite ($Al(OH)_3$) or amorphous aluminium compounds. Kaolinite could then form without meteoric water flushing since such reactions do not produce any other cations like K^+ which would have to be removed. In the North Sea basin abundant authigenic kaolinite is found in the shallowest reservoirs (1.5–2 km) where there is very little or no quartz cement in sandstones and this is the best evidence that most of the kaolinite formed early at shallow depth. The fact that kaolinite is much more abundant in shallow marine and deltaic sandstones than those deposited on deeper submarine slopes is also evidence that kaolinite forms at shallow depth.

4.7 Consequences for Reservoir Quality

Meteoric water flushing dissolves feldspar and mica and precipitates authigenic clay minerals, most commonly kaolinite. This dissolution produces holes which are secondary pore spaces (secondary porosity) but the precipitation of clay minerals like kaolinite reduces the porosity, so that there is little net gain in pore space. Authigenic kaolinite tends to occur as pore-filling minerals and this reduces the permeability. Clean well-sorted sand may increase its specific surface and pore size distribution due to the authigenic kaolinite. The smaller pores (<0.005 mm) in between the authigenic kaolinite crystals may be too small to be filled with oil because of the high capillary entry pressure necessary to infiltrate these pores. The total water saturation will consequently then be higher in the reservoir rock.

Authigenic kaolinite usually occurs as clusters and is rarely pervasive through the sandstones, allowing oil to flow between and around the most densely kaolinite-cemented pores. However, if the kaolinite is altered to illite at greater depth, the damage to the reservoir may be much more severe, due to permeability reduction.

4.8 Mechanical Compaction of Loose Sand

During the first part of its burial history (0–2 km) well-sorted sand is generally still loose if it is not carbonate-cemented. Mechanical compaction may nevertheless be very significant. Experimental compaction of loose sand with an initial porosity 40–42% shows that, depending on grain strength and grain size, the porosity may be reduced to 35–25% at stresses of 20–30 MPa corresponding to 2–3 km of burial for normally pressured rocks (Fig. 4.5). The experimental data show that well sorted coarse-grained sand is more compressible than fine-grained sand (Chuhan et al.

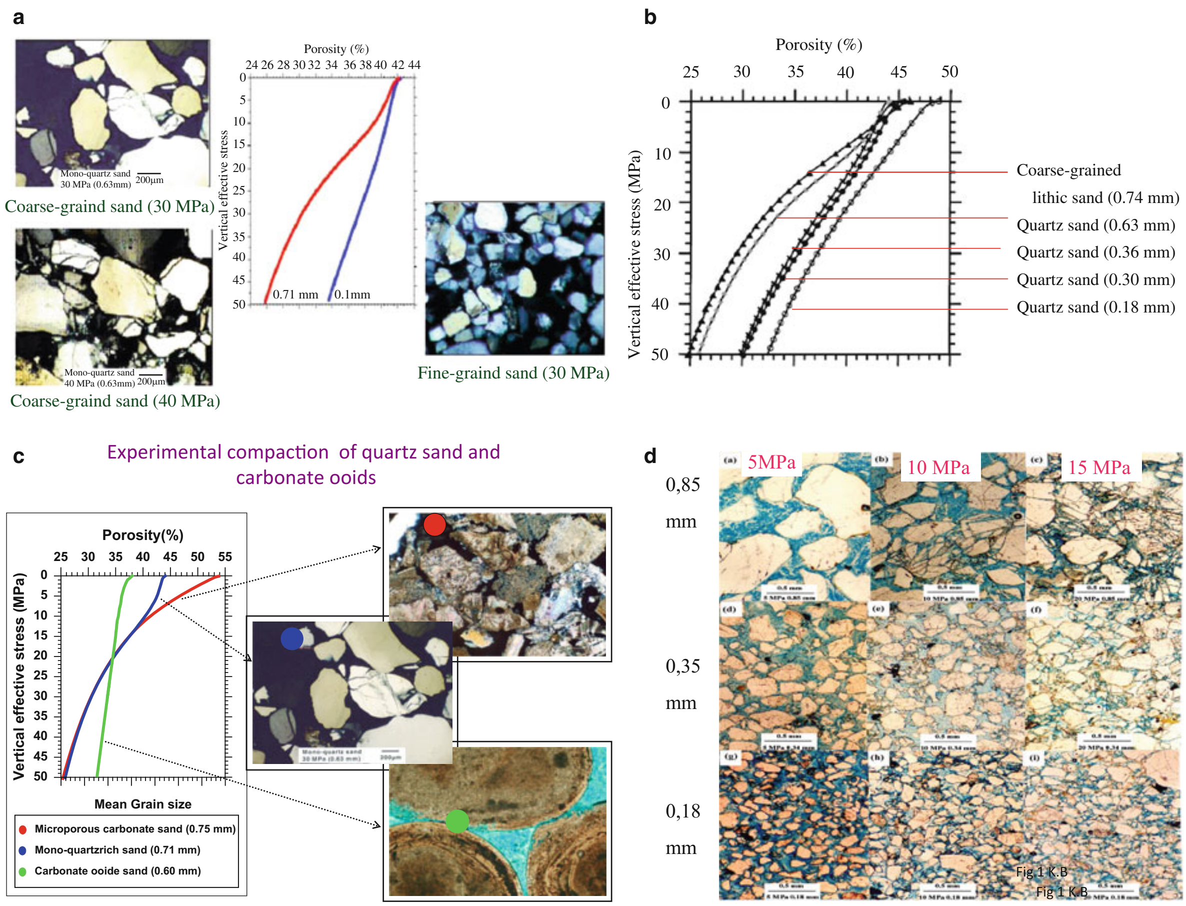

Fig. 4.5 (**a**) Experimental compaction of fine-grained and coarse-grained sand showing that well sorted fine-grained sand is less compressible than coarse-grained sand. (**b**) The porosity loss as a function of grain size due to more grain crushing (from Chuhan et al. 2007). (**c**) Experimental compaction of bioclastic carbonate, quartz grains and ooids. The bioclastic carbonate has a high initial porosity but the compaction curve is similar to that of quartz grains once stresses exceed 10 MPa (from Chuhan et al. 2003). Ooids are soft and rounded so that the grain contacts become larger, reducing the grain-to-grain stress. The grain fracturing and the mechanical compaction is then reduced. (**d**) Experimental shear tests on sand applying different normal stress loading. Coarse-grained sand fractures at lower stresses than fine-grained sand. For medium sand (0.35mm) the horizontal stress must exceed about 15 MPa, corresponding to about 1.5 km burial, before significant grain crushing occurs along faults. Chuhan et al. 2003

2002, 2003). This is because the stres per gain contact is higher in coarse-grained sand than in fine-grained sand. Coarse sand grains have usually an irregular surface so that the contact area is small and the effective stress high.

Most sandstones are however fine grained or medium grained and the frequency of mechanically crushed grains is then not so high before burial to about 2 km (20 MPa). Overpressure reduces the effective stress and will then preserve porosity due to reduced mechanical compaction.

In sedimentary basins with normal geothermal gradients, quartz cementation will stabilise the grain framework and prevent further mechanical compaction at about 2 km burial depth (70–80°C) (Fig. 4.6). At greater depth, compaction is not primarily a function of effective stress but of temperature. In cold sedimentary basins (low geothermal gradients) quartz cementation may not start before 4–6 km burial depth and porosity loss will then occur by mechanical compaction and severe grain crushing up to about 50 MPa effective stress.

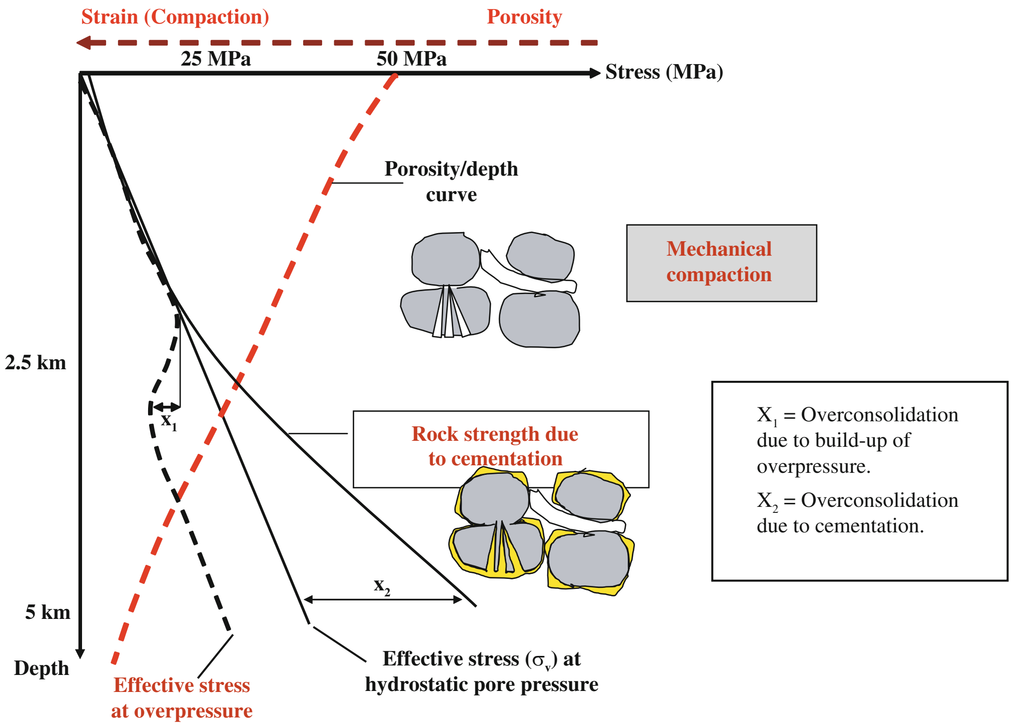

Fig. 4.6 Before sandstones become cemented (at 80–100°C) they compact mechanically as a function of effective stress (depth) by grain reorientation and grain breakage. Relatively small amounts of quartz cement (2–4%?) make the sandstone stiffer and "overconsolidated" so that there is little mechanical compaction (strain) at greater depth (higher stresses)

The degree of porosity loss by mechanical compaction determines the intergranular volume (IGV) at the onset of chemical compaction (quartz cementation).

The IGV measured in some North Sea sandstones varies from about 38 to 28% (Walderhaug 1996) and the net porosity after precipitation of 10% quartz cement will then be very different.

4.9 Sandstone Reservoirs Buried to Intermediate Depth (2.0–3.5 km, 50–120°C)

In basins where there has been mostly continuous subsidence, reservoirs buried to depths shallower than about 2.0–2.5 km are still loose or only poorly cemented, except where there is carbonate cement or high geothermal gradients. This is well documented in most of the North Sea basin and parts of the Gulf Coast Basin (Sharp and McBride 1989, Bjørlykke et al. 1992). In the Statfjord Field where the Middle Jurassic Brent sandstone is buried to 2.5–3 km, there are intervals that are so poorly cemented that it is difficult to obtain good cores because the sandstone disintegrates in the core barrel. Loose sand grains may then be produced with the oil (sand production).

Prior to quartz cementation or other types of cementation the sand grains compact mechanically by sliding and reorientation. Sand grains may also fracture under the overburden stress; coarse-grained sand compacts more due to grain crushing than well sorted fine-grained sand (Fig. 4.5a,b). In basins like the North Sea the quartz cementation increases the rock strength at 2–3 km burial depth (80–100°C) but coarse-grained sand may additionally show significant compaction due to grain fracturing. Compaction experiments show however that carbonate grains (ooids) are less compressible than quartz grains. This is because the grain contacts between carbonate grains will be enlarged due to mechanical and chemical deformation. The surface of quartz grains is irregular and so the area of grain contact is very small, also for large quartz grains, resulting in very high stresses which can cause grain fracturing. Grain fracturing due to shear stress is also sensitive to grain size

(Chuhan et al. 2003) (Fig 4.5d). During faulting coarse grained sand may produce grain crushing at about 1 km depth (10 MP effective normal stress). Fine grained sand requires higher stresses (>15 MPa).

Poorly sorted sandstones and sand with rock (lithic) fragments lose much of their porosity at rather shallow (1–2 km) depth. Quartz cementation strengthens the rocks at a faster rate than the increase in vertical stress from the overburden. Only 2–4% quartz cement will in most cases effectively shut down mechanical compaction in sandstones, so that further compaction is mainly chemically controlled by the rate of mineral dissolution and precipitation (Fig. 4.6). The mechanical compaction is important because it determines the intergranular volume (IGV) which is the porosity prior to quartz cementation. This is typically 25–30% or even more for well sorted quartz-rich sandstones. Sand with even relatively small amounts of detrital clay will compact more than clean sand. Lithic (rock) fragments also compact more readily (Pittman and Larese 1989) and this is reflected in their lower IGV.

Generally, in relatively well-sorted quartz arenites and feldspathic sandstones the porosity is to a large extent destroyed by quartz cementation (Fig. 4.7a).

The amount of quartz cement is mainly a function of the grain surfaces available for quartz precipitation and the time-temperature integral (Walderhaug 1994). High geothermal gradients and slow subsidence rates will therefore tend to increase the amount of quartz cement at a specific depth.

In several of the Upper Jurassic reservoir rocks from the North Sea, amorphous silica from *Rhaxella* sponges dissolved to produce high supersaturation of silica relative to quartz. This caused precipitation of a coating of minute quartz crystals on the surface of clastic quartz grains (Fig. 4.7b, c). This coating of microquartz has prevented or retarded the precipitation of later quartz cement and is the main reason for high porosity and good reservoir quality at great depth (up to 5 km) in these reservoir rocks. The microquartz precipitated at low temperature (60–65°C), when the porewater was highly supersaturated with respect to quartz through the dissolution of opal A or opal CT and while the quartz growth rate was low. At higher temperatures when unstable silicates like opal A, opal CT and smectite have dissolved, the porewater will only be slightly supersaturated with respect to quartz, insufficient to precipitate quartz on the microquartz surfaces which requires higher supersaturation than normal quartz (Aase et al. 1996). *Rhaxella* had not evolved before the Upper Jurassic and so older sandstones like the Middle Jurassic Brent sandstones do not have this type of microquartz.

At temperatures above about 100–120°C some of what we have called kaolinite has recystallised to dickite, which has the same chemical composition. Dickite often occurs as slightly thicker crystals and can also be distinguished from kaolinite on XRD scans. Kaolin or kandite may be used as a common name for these clay minerals.

Carbonate-cemented intervals may be effective barriers to fluid flow. This can be detrimental to the reservoir quality, though in some cases may be useful if they are laterally extensive. Such low permeability layers may then prevent the flow of gas from below the oil/water contact and from above the gas/oil contact into an oil-producing well. This is called coning.

The replacement of K-feldspar or plagioclase by albite is often observed in sandstone buried to about 3 km or more and is referred to as *albitisation*. Albite becomes more stable than K-feldspar because Na^+ is normally the dominant cation in the porewater while the potassium concentration is reduced due to removal by the clay mineral reactions. Albitisation is normally observed as a partial replacement of K-feldspar or plagioclase grains, which does not change the reservoir properties very much. Albitisation of plagioclase will, however, release some Ca^{2+} that may then precipitate as calcite (Boles 1982), though the amount is rather limited.

Smectite may be present in some muddy, and particularly volcanic, sandstones which have been flushed with limited amounts of meteoric water. At temperatures from about 70 to 80°C smectite dissolves and is replaced by mixed-layer minerals and illite. Sandstones containing smectite normally have poor reservoir quality.

Dissolution of smectite and precipitation of illite and quartz will cause a sharp increase in the seismic velocity and rock density and this mineral transition may therefore show up as a horizontal reflector on seismic and could be mistaken for a fluid contact i.e. gas/oil or oil/water contact (Thyberg et al. 2010).

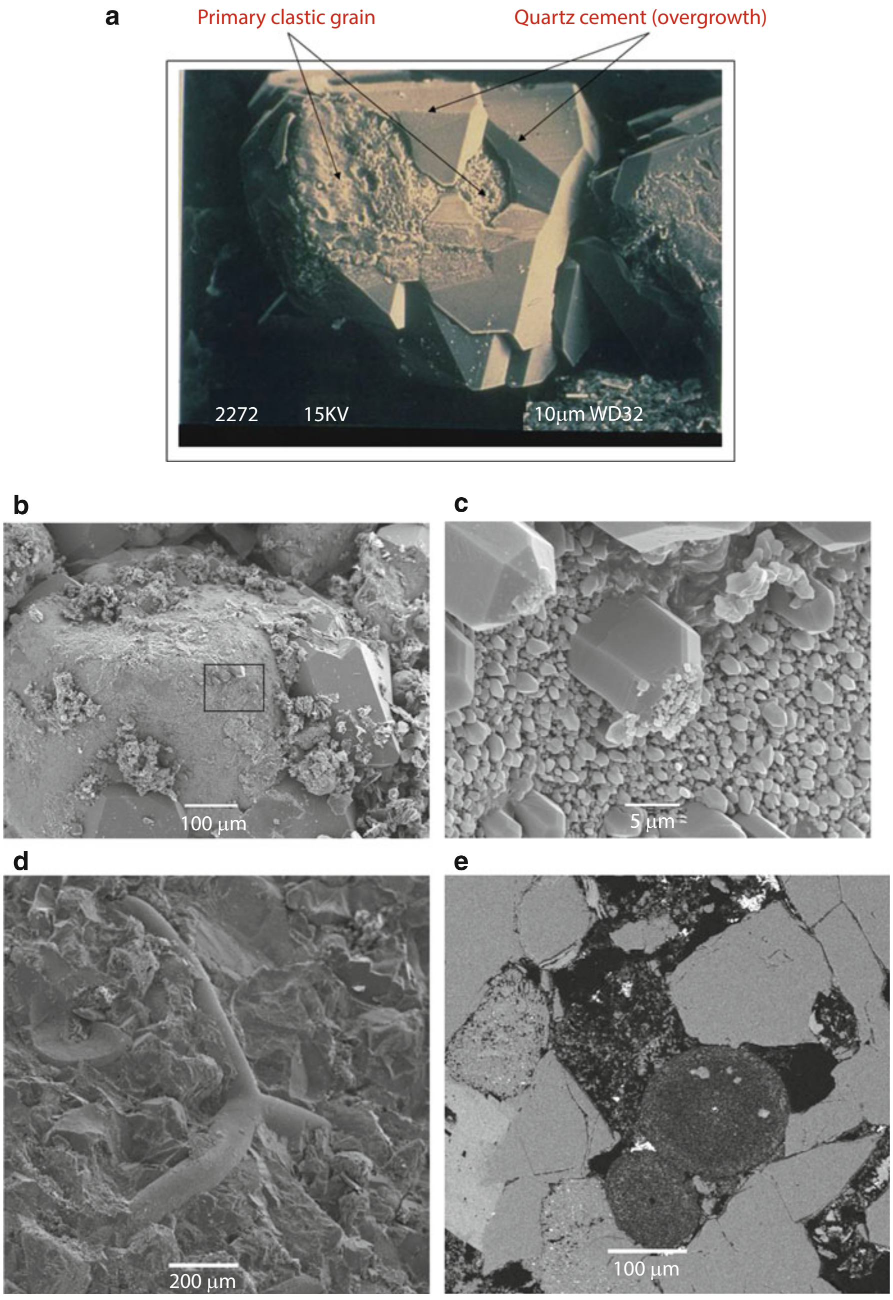

Fig. 4.7 (**a**) SEM image of quartz cement which has grown on detrital quartz grains into the pore space. Note that the authigenic quartz crystals have smooth crystal surfaces while the primary quartz grain has an irregular abraded surface from weathering and transport of the sand grain. Scale bar 10 μm (0.01 mm). Oil inclusions may be found in between the primary grains and the cement, and also in the cement. Sandstone from the Brent Group, North Sea. (**b**) A thin layer of small quartz crystals covering clastic quartz grains has been precipitated at high silica supersaturatation caused by dissolution of organic silica (sponge spicules). (**c**) is an enlargement of Fig. 4.4b showing microquartz crystals that are not overgrown at greater depth. From T.E. Maast (unpublished). (**d**) Pictures (SEM) of the siliceous sponge (*Rhaxella*) which is the source of much of the early quartz cement. To the *left* a broken surface. The sponges have a *circular* outline in cross-section (*thin section*), see Fig. (**e**). Upper Jurassic sandstone from the North Sea (from T.E. Maast unpublished)

4.10 Deeply Buried Sandstones (>3.5–4 km, >120°C)

Once quartz cementation has started and quartz overgrowth has formed, quartz cementation does not stop until nearly all the porosity is lost, unless the temperature drops below 70–80°C.

In most sedimentary basins we find there is a rather strong reduction in porosity and permeability in sandstone reservoirs from about 3–3.5 km to 4–4.5 km burial depth, corresponding to a temperature range from about 120 to 160°C. This is due in most cases to precipitation of quartz cement and diagenetic illite. The rate of quartz cementation increases as an exponential function of temperature and we estimate that the rate may increase by a factor of 1.7 for every 10°C temperature increase (Walderhaug 1996). The precipitation of quartz is also a function of the surface area available for quartz cementation and as quartz cement is filling the pores the surface area available for further quartz growth decreases (Fig. 4.8). The crystal surfaces of quartz have different solubilities and potential for crystal growth and thereby for porosity reduction (Lander et al. 2008).

The temperature history of the sandstones at this stage becomes rather critical in terms of modelling and predicting the amount of quartz cement and remaining porosity. Between 100 and 140°C the rate of quartz precipitation may double four times, i.e. increase 16-fold. By contrast the effective stress increases linearly with depth (under hydrostatic conditions) and thus increases only by 30–40% through the interval from 3 to 4 km of burial. Temperature is therefore by far the main factor controlling the rate of quartz precipitation. This allows the amount of quartz cement and the porosity to be modelled as an exponential function (Arrhenius equation) of the temperature integrated over time and proportional with the surface area available for quartz precipitation. Commercially available programs (Exemplar, Touchstone) have been developed for this purpose based on Walderhaug (1996). Dissolution at grain contacts requires stress, to maintain the contact, so the process is often called "pressure solution", but in the case of silicate minerals temperature is the most important factor. The minerals at the grain contact are also important and dissolution is enhanced by mica or clay minerals at the grains contacts (Bjørkum 1996). Contacts between mica or illitic clay and quartz are preferred areas of dissolution. The rate-limiting process in quartz cementation seems to be the rate of nucleation and precipitation in the pore space and the reaction is then surface-controlled (Bjørkum et al. 1998). Quartz cementation is therefore insensitive to variations in effective stress in the grain framework. If the dissolution process had been rate-limiting, quartz cementation would have been more sensitive to the effective stress in addition to temperature and surface properties of the minerals. The silica dissolved at grain contacts or along stylolites is transported by diffusion to the grain surfaces where the quartz overgrowth forms (Oelkers et al. 1996) (Fig. 4.9).

If the transport of silica was the rate-limiting factor for quartz cementation (transport-controlled) we would expect to observe a concentration of quartz cement close to stylolites of thin clay laminae, where dissolution occurs and where the concentration in the porewater would be highest. When dissolution is concentrated along stylolites with relatively large spacing (>50–100 cm), studies suggest that the amount of quartz cement may decrease away from the stylolites, indicating some degree of transport control (Walderhaug and Bjørkum 2003). The stylolites represent a barrier for fluid flow both during

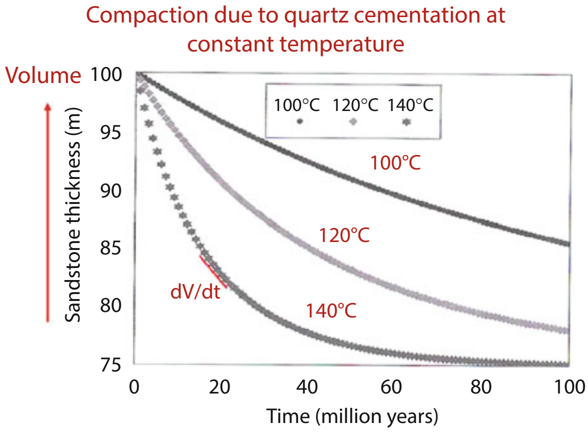

Fig. 4.8 Modelling of quartz cementation and chemical compaction due to quartz dissolution and cementation as a function of time and temperature (from Walderhaug et al. 2001). We see that the rate of porosity loss (compaction, *dv*/*dt*) is highest at high temperatures and also when the porosity is still relatively high. When the porosity is reduced the surface area available for quartz cementation becomes smaller so that the rate of cementation slows down

compaction and during production, but may not be completely continuous laterally.

We have seen that the rate of quartz cementation is a function of temperature, the degree of supersaturation and the surface area available for quartz cementation. A consequence of a surface-controlled quartz precipitation model is that quartz cementation will continue as long as the temperature is above the threshold temperature for quartz growth (70–80°C) and there is remaining porosity in the sandstone. It is important to note that quartz cementation, and hence sandstone compaction, will continue also during basin inversion and uplift, but at a slower rate (Fig. 4.10). When uplifted to shallower depths (temperature <70–80°C) the sandstone is unloaded and without any chemical compaction there will be net extension due to elastic expansion due to reduced stress.

As long as the temperature is higher than 70–80°C the cementation process will proceed but at a slower rate, modifying the normal porosity/depth relation found elsewhere in the basin. During progressive burial quartz cementation must continue until all available porosity is filled and the sandstone becomes a well-cemented hard quartzite after exposure to 200–300°C for several million years.

If the surfaces of sand grains are coated with other minerals, or with substances like petroleum or bitumen, quartz overgrowth is hindered or at least stopped for some time (Fig. 4.9). A thin layer of authigenic chlorite has proven to be very effective in preventing quartz overgrowth. This has been described from many places around the world like the Tuscaloosa

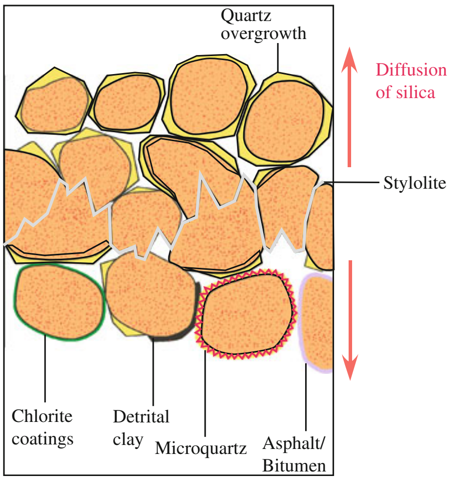

Fig. 4.9 Schematic illustration of a stylolite. The dissolved silica is transported away from the clay-rich stylolite by diffusion. This makes more long distance and advective transport of silica difficult. The rate of precipitation of quartz cement is a function of the surface area available. Grain coatings such as chlorite, illite, detrital clay, iron oxide (haematite), microquartz and bitumen prevent or retard quartz cementation

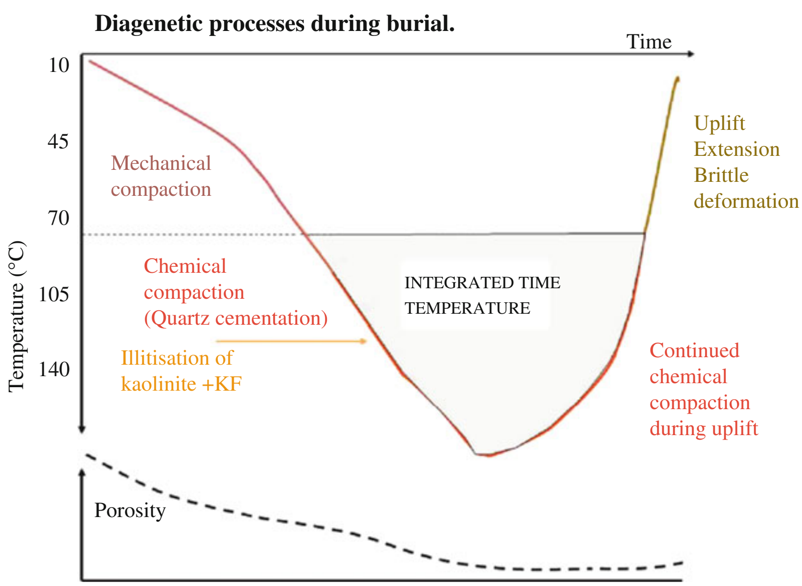

Fig. 4.10 Diagenetic processes, mainly quartz cementation, as a function of temperature and time. Note that quartz cementation will continue also during uplift as long as the temperature exceeds 70–80°C

sandstones of Southern Louisiana (Pittman and Larese 1992).

Sandstones with grain coatings may remain uncemented down to 4–5 km burial depth and be subjected to 40–50 MPa effective stress, causing pervasive grain crushing (Chuhan et al. 2002).

Under the microscope, and particularly when using cathodoluminescence, we can see that some of the sand grains have been fractured and later healed by quartz cement (Fig. 4.11a, b). The fractures in the clastic grains do not usually continue through the quartz overgrowth, demonstrating that they predate the quartz cementation.

At Haltenbanken, offshore mid-Norway, the Jurassic Tilje and Garn formations have abundant chlorite coatings and also illite coatings (Ehrenberg 1993, Storvoll et al. 2002) (Fig. 4.12).

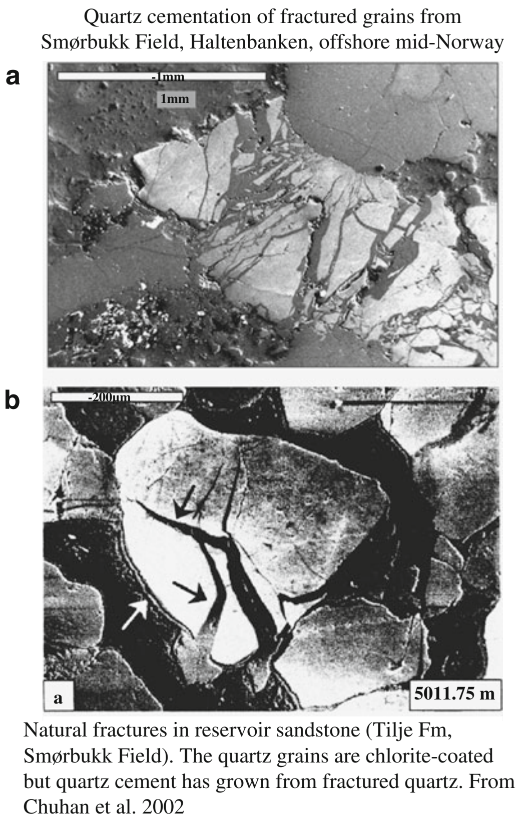

Fig. 4.11 (**a**, **b**) Cathodoluminescence pictures showing that quartz grains have been subjected to fracturing prior to quartz cementation. Quartz cementation was delayed by chlorite coatings but grain fracturing exposed fresh quartz surfaces from which quartz could grow into the pore space. Authigenic quartz is darker than the clastic high temperature quartz

This is the main reason for the good reservoir properties in petroleum discoveries on Haltenbanken, where the porosity sometimes exceeds 25% at more than 5 km depth. The Kristin Field in Haltenbanken is one example of reservoirs that have preserved relatively good reservoir quality at great depths (>5 km). Reservoirs characterised by high temperatures and high pressure (>150°C (300°F)) and 89 MPa (10,000 psi). are often referred to as HTHP reservoirs. They may represent technical challenges with respect to drilling and production because of the high fluid pressures and the potential for gas leakage.

At temperatures of about 150°C most sandstone reservoirs have too little porosity due to quartz cementation. The preservation of porosity in these HTHP reservoirs depends on clay coatings on quartz grains retarding precipitation of quartz cement, as is the case in the Kristin Field. If the reservoir had been exposed to high temperature for a relatively short geologic time the porosity would be higher due to less quartz cement.

To predict the distribution of such high porosities at great depth we need to understand what controls the development of authigenic chlorite. The chlorite found in marine sandstones is most probably an alteration product of an earlier iron silicate phase (precursor) formed on the seafloor and which may be linked to facies. This may be iron- and magnesium-rich smectites coating the primary quartz grains. Illite may also form an effective coating preventing quartz overgrowth (Heald and Larese 1974, Storvoll 2003) and this may have formed from smectite.

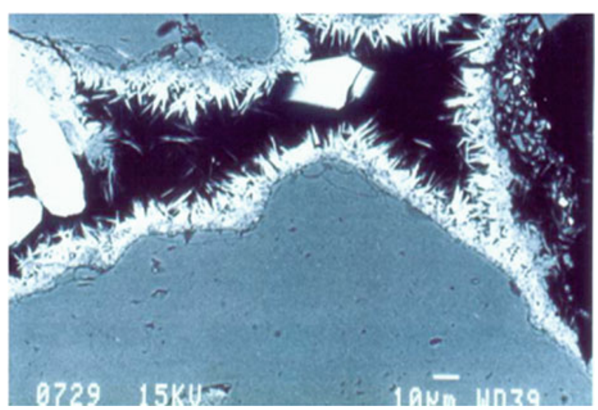

Fig. 4.12 Quartz grains coated with chlorite preventing quartz overgrowth. Note pore-filling authigenic kaolinite. Tilje Formation, Haltenbanken

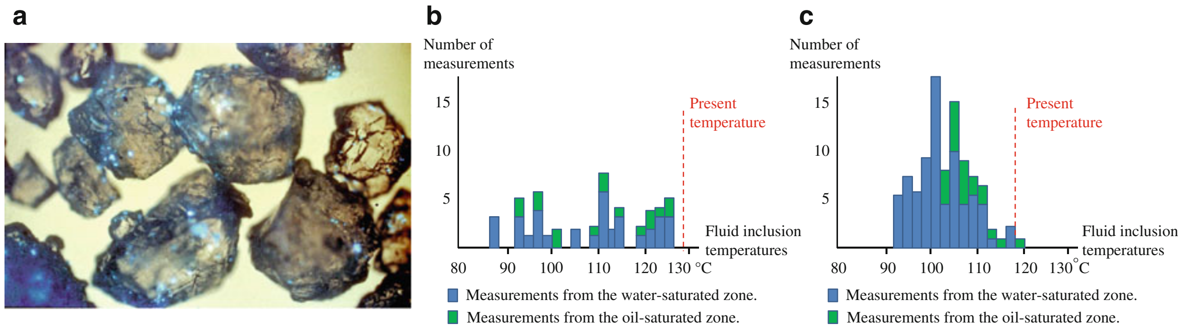

Fig. 4.13 (**a**) Inclusions of oil in quartz cement on sand grains. (**b**) Temperatures from oil inclusions in the Fulmar Field, North Sea (Saigal et al. 1992). (**c**) Fluid inclusion temperatures from Jurassic sandstones, Haltenbanken (offshore mid-Norway) (Walderhaug et al. 1990b)

It is typical that clean well-sorted sandstone has the best porosity down to about 3.5–4 km but then loses its porosity rapidly due to quartz cementation. Sandstones with a moderate clay content lose more porosity during mechanical compaction compared to the clean sandstone, but may preserve more porosity at greater depth due to retarded quartz overgrowth.

4.11 Fluid Inclusions in Quartz Cement

Inclusions of small drops of oil may be trapped in quartz cement and show up well in fluorescent light (Fig. 4.12a). Fluid inclusion data from quartz helps to constrain the temperature for quartz cementation and in sandstone reservoirs it has been demonstrated quite clearly that quartz cementation continues after oil emplacements in a reservoir (Karlsen et al. 1993, Walderhaug 1990).

The lowest fluid inclusion temperatures in quartz cement indicate an onset of quartz cementation close to 70–80°C (Burley et al. 1989, Walderhaug 1994a). In the Ula Field (North Sea basin), fluid inclusion temperatures in quartz cement range from 86°C to 126°C (Fig. 4.13b), the highest temperature being close to the present day reservoir temperature which is also the maximum burial depth and temperature (Saigal et al. 1992). Fluid inclusions from 24 samples from 11 different reservoir units from the North Sea and Haltenbanken also show temperatures from about 80°C to values close to the present reservoir temperatures (Walderhaug 1994a). This shows that quartz cementation occurs as a continuous process at a rate controlled by the temperature (Walderhaug 1994b) (Fig. 4.13c).

There is no evidence that quartz cementation is episodic, controlled by the supply of silica, or that quartz cementation stops after the sandstones have become oil-saturated (Walderhaug 1990, Saigal et al. 1992) (Fig. 4.13). In a water-wet reservoir precipitation can still continue in the remaining water around the grains. At high oil saturation, the transport of silica by advection as well as by diffusion becomes much less efficient. The continued growth of quartz cements after oil emplacement results from the closed system nature of quartz cementation in sandstones.

In an oil-wet system, however, quartz can not precipitate on the grain surfaces and oil or bitumen may become very effective coatings. Asphaltic oil and bitumen formed by biodegradation or other processes may preserve good reservoir quality, particularly if the heavy oil only occurs as a grain coating.

In summary, quartz cementation is controlled by the slow kinetics (high activation energy) for quartz cementation and normally a minimum temperature of 70–80°C is required. This is however also dependent on the pH. In sedimentary basins marine porewater starts out with a pH up to 7 but quickly becomes more acid due to the build of CO_2 and other reactions with the minerals present. At 3–4 km depth the pH may typically be 4.5–5 but at 120°C the pH is close to neutral (see Chap. 3). The rate of quartz cementation is then lowered by the pH but increased by higher temperatures. At very high pH quartz cementation may occur at the surface and silcrete is fine-grained quartz formed in soils due to concentration of

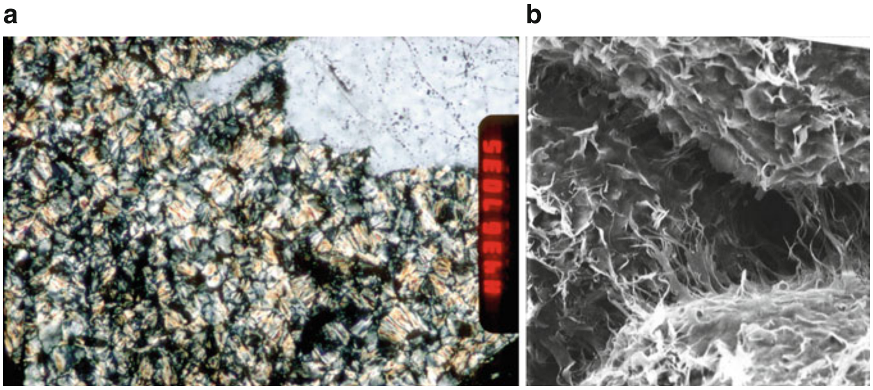

Fig. 4.14 (**a**) Pore-filling illite replacing pore-filling kaolinite preserving the kaolinite textures in Jurassic sandstone, Haltenbanken, offshore mid-Norway. (**b**) Pore-filling illite probably altered from smectite. Triassic sandstone from the North Sea

porewater by evaporation, and quartz is also forming in some alkaline African lakes.

Authigenic illite consists of thin hair- or plate-like minerals and it is fairly obvious that they would have a detrimental effect on reservoir quality by reducing the permeability (Fig. 4.14a, b). SEM images are routinely taken from dried-out cores where the illite appearance is no longer representative of its morphology in the reservoir. When cores are dried without destroying the delicate illite morphology the pore space often looks like it has been filled with rockwool. Illite can often be seen to grow at the expense of kaolinite and may also form by alteration of smectite. Although authigenic illite may also be observed on fractures and other places where there are no obvious precursor minerals, it is most commonly found as a replacement of an earlier Al-rich mineral phase.

Because of the low Al-solubility in porewater, illite will in most cases precipitate where the source of Al is available locally from a dissolving mineral. Calculations suggest that the solubility of aluminium is only about 1 ppm at 150°C and that organic acids have little effect in terms of increasing its solubility (Bjørlykke and Aagaard 1992).

The formation of illite from smectite via mixed-layered minerals is well known and occurs in sandstones in the temperature range of 70–100°C. Sandstones with abundant smectite are poor reservoir rocks at the outset, and illitisation of such rocks may itself slightly improve reservoir quality as illite has a lower specific surface area than smectite. In better-sorted and potentially good reservoir rocks kaolin minerals (kaolinite or dickite) are the most important precursors for illite. However, the formation of illite requires potassium, and K-feldspar is usually the only significant source present in the sediment.

$$\mathrm{KAlSi_3O_8 + Al_2Si_20_5(OH)_4 = KAl_3Si_3O_{10}(OH)_2 + 2SiO_2 + H_2O}$$

$$\text{K-feldspar} + \text{Kaolinite} = \text{Illite} + \text{Quartz}$$

High temperature detrital K-feldspar contains some sodium and the illite formed contains less potassium than indicated here, where the formula for muscovite is used.

The reaction between K-feldspar and kaolinite occurs at about 130°C and above this temperature these two minerals are no longer thermodynamically stable together. In the North Sea basin and at Haltenbanken this corresponds to a burial depth of about 3.7–4 km. A sharp increase in the illite content in sandstone reservoirs is observed at this present day depth (Bjørlykke et al. 1986, Ehrenberg 1992). From thin sections and SEM images the illite can be seen to replace kaolinite, and potassium then has to diffuse from the K-feldspar to the kaolin. If the matrix is well-cemented the rate of diffusion is reduced, and the minerals are then able to co-exist at higher temperatures if they do not occur close together.

In sandstones with little or no K-feldspar, however, kaolin remains stable at greater depth as it is not dissolved and replaced by illite. The formation of illite can therefore be predicted from the sandstone provenance with respect to K-feldspar supply, and from the early diagenesis and freshwater flushing with respect to the distribution of kaolinite. If a sandstone is derived from an albite-rich gneiss the K-feldspar content is likely to be too low and much of the kaolinite would then not be illitised.

Similarly, not much illite will be formed in sandstones with little kaolinite or smectite. Both in Haltenbanken and the North Sea there are Jurassic reservoirs where plagioclase is the dominant feldspar and where the low K-feldspar content is unable to supply the necessary potassium for illitisation of kaolinite. The low illite content in such reservoirs preserves better permeability. This is a direct function of the provenance and could be due to erosion of albite gneisses rather than granitic gneisses.

The distribution of authigenic illite in sedimentary basins like the North Sea and Haltenbanken shows that illite formation is strongly controlled by the present day burial depth and temperature. The increase in illite content at about 3.7–4.0 km is usually very sharp, indicating a temperature-controlled reaction rather than a high kinetic reaction rate when the association of kaolinite and K-feldspar becomes thermodynamically unstable. Basin loading from thick Pleistocene sequences in these areas suggests that the illite formed recently.

K-Ar dating of illite gives variable ages for the formation of illite. This is probably because even very small amounts of detrital (older) mica or feldspar will produce too-old ages (Hamilton et al. 1989).

4.12 Porewater

In a sedimentary basin the amount of solids dissolved in porewater is very small compared to the volume of solids. During burial diagenesis, significant precipitation of authigenic minerals must be accompanied by dissolution of other minerals or the same mineral, as in the case of pressure solution. Even if the porewater is supersaturated with respect to a certain mineral, only very small amounts can precipitate before the porewater attains equilibrium. Precipitation of new minerals requires that other minerals dissolve because the porewater has very little capacity to store ions. This has been quantified using chemical modelling (Giles 1997) from which it can be concluded that extremely large fluid fluxes are required through the pores in order to dissolve or precipitate significant amounts of cements like calcite or quartz.

The porewater reacts with the minerals it is in contact with, and with increasing temperature the composition becomes more and more in equilibrium with these minerals. This is because the kinetics of the mineral reactions become faster. The porewater in sedimentary basins consists of solutions buffered by the minerals present. The pH is controlled partly by the carbonate reactions (pCO_2), and the pH will decrease from 7.5–8 in the seawater to 4–5 at 3–4 km depth. As temperature rises above 100°C, silicate reactions become increasingly important, e.g. between K-feldspar, kaolinite and illite which will determine the K^+/H^+ ratio.

Organic acids are weak acids which can not significantly change the pH in the strongly buffered porewater. The buffering capacity of organic acids has been shown to be orders of magnitude lower than for the carbonate and silicate systems (Hutcheon and Abercrombie 1990). Organic acids generated in source rocks like the Kimmeridge Clay Fm in the North Sea are likely to be neutralised by reactions with calcite which is commonly present in these source rocks. The limited effect of organic acids and CO_2 on mineral dissolution and diagenesis has also been shown experimentally (Barth and Bjørlykke 1993).

When porewater moves it will nearly always cause some dissolution and precipitation but this is rarely significant due to low velocities. The volume of minerals dissolved or precipitated can be calculated using the following equation:

$$V_c = F\,t \sin\alpha\,(dT/dZ)\,\Delta S/\rho$$

The volume of precipitated mineral V_c is a product of the fluid flux integrated over time (t), the angle of fluid flow relative to the isotherms (α), the geothermal gradients (dT/dZ), the solubility as a function of the temperature (ΔS) and the density of the mineral (ρ).

The solubility gradient (ΔS) is 1–3 ppm/°C, depending on the temperature (Wood 1986). This means that at 100–150°C about 2 ppm of quartz precipitates for each degree the porewater is cooled. This gives a solubility gradent of 2.10^{-6}/°C. If the geothermal gradient is $30°C/km\,(3 \times 10^{-2}°C/m)$, porewater must move upwards more than 30 m to reduce the temperature by 1°C and the quartz cement is distributed through these 30 m of sandstone. Assuming vertical flow ($\sin\alpha = 1$) and geothermal gradients close to 30°C/km (3.10^{-2}°C/m), a solubility gradient of 2.10^{-6}/°C and a quartz density of 2.65 g/cm^3 we obtain:

$$Vc = F \cdot t \cdot 2.3 \times 10^{-8}$$

From the above equation we see that each 1 m^3/m^2 will precipitate about 2.3×10^{-8}m^3 of quartz. To precipitate 10% quartz cement ($Vc = 0.1$) requires a total flow (integrated flux over time)) of ($F \cdot t$) of about $4 \cdot 10^6$ m^3/m^2. This assumes that the porewater is in equilibrium with the mineral phases which is true at depth with temperatures exceeding 80–100°C. At 30% porosity a water column of 1,200 km must pass through a sandstone layer to introduce 10% quartz cement.

This is clearly impossible in sedimentary basins. In addition compaction-driven porewater is not flowing upwards in relation to the surface and is therefore normally not subjected to cooling which would cause precipitation of quartz (see Chap. 10).

At shallow depth the temperature is low and the fluid flow rate high, so in the zone of meteoric water flushing this may not be true. The porewater may then be undersaturated or supersaturated, particularly with respect to silicate minerals. When temperatures exceed 100°C the porewater will approach equilibrium with the minerals, both because of higher reaction rates and low flow rates.

Small amounts of calcite are nearly always present, at least in marine sediments. Calcite has a retrograde solubility meaning that the solubility normally decreases with increasing temperature. The solubility also depends on the pressure, but in most cases it is the temperature effect which is strongest. Upwards (cooling) porewater flow, which should precipitate quartz, will dissolve calcite at a rate which is 30–100 times faster (Bjørlykke and Egeberg 1993). We may therefore conclude that if calcite was present in a sandstone very little quartz could have precipitated until all the calcite had been dissolved.

Thermal convection is probably not very significant in sedimentary basins except where there are hydrothermal heat sources (Bjørlykke et al. 1988). If thermal convection did occur at a significant rate, however, quartz could precipitate because the same water could be used over again, precipitating quartz and dissolving calcite on the way up, and dissolving quartz and precipitating calcite on the way down when the porewater is heated. All the calcite would then be dissolved and quartz would be precipitated by this process.

4.13 Effect of Oil Emplacement

When oil migrates into a reservoir rock the water content is reduced to a percentage of the porosity corresponding to ''irreducible water saturation' if the rock is water-wet. This may vary from 10% water content in clean sand to 50% or more in clay-rich sandstone, the value depending on the amount and type of clay present. The traditional assumption has been that the emplacement of oil stops, or at least slows down, the rate of diagenetic processes and hence the rate of porosity reduction.

If the transport of silica by either diffusion or advection was rate-limiting for the quartz cementation one would indeed expect the rate of quartz cementation to be very much reduced. Fluid inclusions in quartz cement, however, clearly demonstrate that in fact quartz continues to grow after oil emplacement in sandstone reservoirs (Bjørkum and Nadeau 1998, Walderhaug 1990). The explanation for this is that silica is transported along the thin film of water between the mineral grains and the oil phase. It is possible that the rate of quartz cementation could be slower after oil emplacement but this would imply that the quartz cementation would no longer be surface controlled, but transport controlled. It is very difficult to prove that a higher porosity in the oil-saturated part of a reservoir is due to the introduction of oil. There are usually so many other variables such as facies that influence the final reservoir porosity. Since the oil is emplaced gradually and the oil/water contact moves downward over time, a sharp difference in porosity should therefore not be expected right at the present OWC if quartz cementation was a function of oil emplacement.

In gas reservoirs the saturation may be rather high in clean sand, perhaps reducing the water film around the grains and the quartz overgrowth, but the degree to which this might apply is uncertain.

Biodegraded and asphaltic oil or bitumen will stick to the grain surfaces and effectively prevent quartz overgrowth but then some of the porosity may be lost to the bitumen and heavy oil.

4.14 Different Types of Sandstone

It is important to distinguish between sandstones of different primary composition:

Volcanoclastic sandstones may vary greatly in composition depending on the volcanic source and the depositional environments. Basic volcanic rocks in particular have a very low content of stable grains like quartz, but a high content of basic feldspar and pyroxenes which break down rapidly, both mechanically and chemically. Matrix-rich sandstones like greywackes may have had a higher sand content at the time of deposition because many of the grains were unstable during diagenesis and became effectively part of the matrix. What were deposited as grains of volcanic rock fragments may be squeezed so that they become a chlorite-rich matrix.

Volcanoclastic sandstones lose most of their porosity at rather shallow depth (<1–2 km) and therefore make poor reservoir rocks. However, the geothermal gradients in volcanic regions may be high, causing source rocks to mature at shallow burial depth and thus increase the potential for migration into shallow structures.

Lithic sandstones have a high content (>10%) of rock fragments. Normal and coarse-grained granites and gneisses produce grains that mostly consist of a single mineral while sandstones derived from finer-grained igneous and metamorphic rocks are mostly comprised of rock fragments. Rock fragments are generally weaker than quartz and feldspar grains, as has been demonstrated experimentally (Pittman and Larese 1991).

Arkoses contain more than 25% feldspar and such sandstones are typical of tectonically active basins like rift basins where the erosion, transport and deposition of basement derived rocks is fast, leaving little time for feldspar to weather. Temperature and rainfall also play a role here. Arkoses compact more mechanically than quartzitic sandstones, leaving a smaller intergranular volume to be cemented with quartz at greater depth. The area available for quartz cementation is also reduced since quartz does not grow on feldspar.

Feldspathic sandstones and quartzites are the most common sandstone reservoir rocks. The feldspar content is usually a function of the source climate and the relief in the drainage area. On tectonically stable cratons sediments are repeatedly eroded and deposited and some feldspar and mica is dissolved during each cycle.

Palaeozoic quartzites typically occur as transgressive sheet sands on cratons. On the North American craton there are good examples of this in the Lower Palaeozoic sequence. Such clean shallow marine sandstones have extremely good reservoir properties at shallow to moderate burial depth. This is likewise the case with aeolian sandstones. Fluvial sandstones are also normally well-sorted in such environments because they are often reworked aeolian sands.

Carbonate cement in shallow marine sandstones is mostly derived by recrystallisation of calcareous organisms. Meteoric water will dissolve aragonite and precipitate calcite in sandstones, producing early cement.

In modern environments, particularly in beach and shoreface settings, fragments of crushed calcareous organisms are quite common. We find less carbonate cement in fluvial sandstones because of the lower biogenic carbonate production in freshwater. Carbonate cement has a local source in most cases, but may be redistributed and concentrated by diffusion. The range of effective diffusion is generally small (<1 m) because the porewater is in equilibrium with calcite and there are small concentration gradients. Advective flow will transport dissolved carbonate but can not precipitate tight carbonate cement because the permeability decreases as precipitation proceeds. The advective flow will then tend to bypass the volume where carbonate cementation has started. In the case of compaction-driven upwards-directed (cooling) porewater, the solubility of calcite will increase, causing dissolution rather than precipitation.

Aeolian sandstones and other desert sandstones generally show less evidence of meteoric water flushing than fluvial and shallow marine sandstones. Sandstones like the Permian Rotliegend from the southern North Sea have relatively low amounts of kaolinite and more smectite or illite as pore-filling cement. However, even deserts have groundwater so some leaching occurs. Fluvial sediments will normally be flushed by groundwater after deposition and in most cases show ample evidence of feldspar leaching. Reworking of such sediments will bring authigenic kaolinite into the clastic clay fraction. Continental sandstones often have haematite or manganese oxide coatings on quartz grains and this may inhibit quartz overgrowth.

Turbidites form important reservoirs in many basins and although the reservoir quality may be less favourable than within shallow marine sandstones, they often form extensive vertically-stacked reservoir sequences and this may compensate for the lower porosity.

Turbiditic sandstones generally have a higher clay content than shallow marine sandstones. There is however a wide range of clay contents in turbidites from rather clean, usually proximal and channel facies, to more clay-rich distal and overbank facies.

Sandstones with clay contents higher than 10–15% lose their porosity rapidly with mechanical compaction, because the detrital clay acts as a lubricant in the compaction of the quartz grains. Turbiditic sandstones of Paleocene and Eocene age form very important reservoirs in the North Sea. The Frigg sandstone in the Frigg Field in the Norwegian Sector is an example, where the reservoir quality is rather good despite representing a distal facies relative to the Shetland Platform where the sand originated.

Because turbidites are deposited further away from land, they are less exposed to flushing by meteoric water. Turbidites normally contain less evidence of feldspar dissolution and authigenic kaolinite precipitation than shallow marine and fluvial sandstones. In proximal turbiditic facies, however, the sandstones may have good contact with the meteoric water lens. In the North Sea, Tertiary turbidites generally have a relatively low content of authigenic kaolinite, but it may be higher in proximal Cretaceous turbidities which formed around islands produced by local uplift.

The total flow of meteoric water through sandstones is a function of meteoric water flux and the sedimentation rate. At low sedimentation rates a given volume of sand spends more time in the shallow zone of meteoric water flushing.

Cretaceous and Tertiary turbidite sandstones are often very tight due to pervasive carbonate cement. This may be due to pelagic carbonate organisms being mixed in with the turbidite sand and recrystallising into carbonate cement. This makes many Tertiary sandstones hard and indurated even if they have not been buried very deeply.

Many of the planktonic carbonate organisms developed during the Jurassic, so before then the sources of carbonate cement in deep sea sandstones were more restricted.

4.15 Predictions of Reservoir Quality

The oil industry has a practical need to be able to predict the properties of reservoir rocks ahead of drilling. When planning petroleum production the rock properties between the wells must also be estimated. Particularly in the deeper reservoirs, porosity is the most important factor determining the economic viability of a prospect. The main diagenetic processes with quartz cementation are summarised in Fig. 4.15.

In a relatively mature basin the porosity/depth functions of the different reservoir rocks can be treated statistically so that the uncertainty of the estimates can be expressed. The estimates based on the statistical averages can also be adjusted up or down as a function of temperature, stress etc., depending on what the interpreter considers most significant. The Middle Jurassic Brent sandstone in the North Sea has been intensively studied and a relatively linear trend found between burial depth and porosity (Giles et al. 1992, Bjørlykke et al. 1992, Ramm et al. 1992, Wilson 1994). Porosity predictions will depend on the primary sediment composition and the subsequent compaction processes.

Above we have discussed some of the processes that cause reductions in porosity and permeability. All these processes are driven towards denser packing of grains and thermodynamically more stable mineral assemblages as the stress and temperature increases during burial. The kinetics of mineral reactions determines the rate of thermodynamic equilibriation, which increases as an exponential function of temperature.

The rate of compaction as a function of stress can be measured experimentally using rock mechanics testing procedures. However, the reactions involved in chemical compaction are so slow, particularly in silicate rocks, that it is difficult to reproduce them in the laboratory although in some cases this is now becoming possible.

In the field of clastic diagenesis, petrographic observations about mineralogy and textural relationships are used to interpret the sequence of dissolution and mineral precipitation and its relationship to changes in porosity and permeability. It is however important to consider the geochemical constraints on diagenetic reactions. During burial the reaction must add up so the dissolution is balanced by precipitation because there are strong limitations with

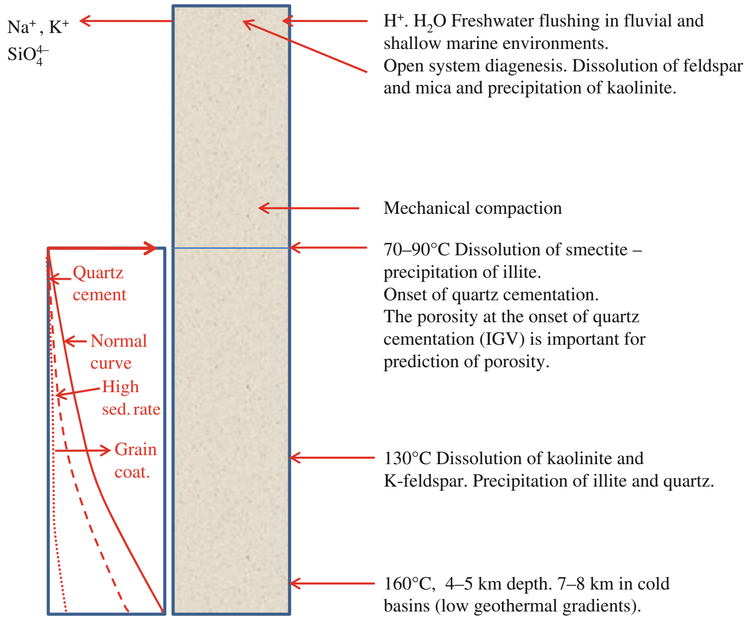

Fig. 4.15 Summary of the most important processes in clastic diagenesis. Dissolution of feldspar and mica requires a through flow of meteoric water removing K^+ (Na^+) and silica before kaolinite can be precipitated. This process is most dominant in fluvial and shallow marine sediments at shallow burial. At greater depth diagenetic reactions are nearly isochemical at a scale of 1–10 m distance. The composition of the dissolved material is equal to what is precipitated. Carbonate cements are usually derived from biogenic carbonate or clastic carbonate grains. Quartz cementation is controlled by temperature (geothermal gradients), subsidence rates and the presence of grain coatings

respect to supply and removal of solids dissolved in the porewater.

Changes in mineralogy or porosity with depth may provide useful depth trends within an area, but in a sedimentary basin the initial mineral composition may vary laterally. This is also the case for early diagenetic processes like meteoric water flushing and marine cementation. The observed changes with depth may therefore also reflect some of these factors and not only the burial depth.

Based on the theory that the rate of quartz cementation is controlled only by temperature, time and the grain surface available for quartz precipitation, the amount of quartz cement and consequently the porosity can be modelled (Bjørkum et al. 1998, Walderhaug 1996). The presence or absence of clay or other coatings is the most critical input for this modelling because it determines the area available for quartz cementation. Prediction of reservoir properties must start from sedimentological facies models. The depositional environment and the provenance of the clastic sediments determine the starting composition for the diagenetic processes.

Early diagenetic processes like marine carbonate cementation and meteoric water flushing are also linked to facies and they strongly influence the burial diagenesis and the porosity reduction at depth. The precursor minerals controlling the growth of chlorite coating are also probably to a large extent controlled by facies. The distribution of silica organisms (like *Rhaxella*) which can produce microquartz coatings is linked to the environment and ecology.

A broad geological background is therefore required to synthesise all the factors that have to be considered before modelling or making semi-quantitative predictions of reservoir quality. The capacity of porewater to keep solids in solution is always rather limited and mineral dissolution and precipitation must therefore balance.

At greater burial depth (3–4 km) the solubility of silicate minerals increases but the volume of porewater is low and the potential for supersaturation and undersaturation strongly reduced. Precipitation of new mineral phases must therefore be linked to the dissolution of other minerals or of the same mineral.

Assuming that the burial diagenetic processes are relatively isochemical, the reservoir properties can be predicted from depositional facies and provenance studies which define the starting composition for burial diagenesis. Both observations and theoretical arguments suggest that advective transport can not significantly change the rock compostion below the reach of meteoric water flow. Short distance transport by diffusion may nevertheless be important.

Quartz cementation was often interpreted to occur as events of relatively short duration (approximately <10 million years) that could start and stop late in a sandstone's burial history (Robinson and Gluyas 1992). This would imply that the quartz cementation was controlled by advective transport of silica in solution and the source of silica from other reactions. Fluid inclusion data, however, shows that quartz cementation occurs throughout the temperature range corresponding to the burial history.

Modelling of quartz precipitation (Walderhaug 1996, Olkers et al. 2000, Walderhaug et al. 2001) is based on the assumption that this is a continuous process controlled by the kinetics and therefore by temperature. This is the basis for the practical models widely adopted by the petroleum industry (e.g. Exemplar and Touchstone).

The assumption that the quartz precipitation is the rate limiting factor may not always be strictly true in clean quartz arenites where silica sources (stylolites) are widely spaced, resulting in decreasing quartz cementation away from the stylolites (Walderhaug and Bjørkum 2003). The modelling does, however, require that the burial curve and the temperature as a function of geological time are known. It is also very sensitive to changes in the primary sediment composition, which strongly influence both the mechanical compaction and the chemical reactions. The porosity loss and increased sediment density resulting from chemical compaction as a function of temperature cause basin subsidence (Bjørkum et al. 1998, 2001; Bjørkum and Nadeau 1998). Temperature-driven chemical compaction results in a volume reduction (shrinkage) and the strain is then independent of effective stress. As a result, differential stresses in siliceous rock will be relaxed by the compaction processes during basin subsidence as long at the temperature exceeds about 80°C (Bjørlykke 2006).

4.16 Porosity/Depth Trends in Sedimentary Basins

Data from wells in sedimentary basins which have undergone almost continuous subsidence can be regarded as records of a natural compaction experiment. We can use the log porosity and in the cored intervals we have data from core plugs.

At depths shallower than about 2 km (80–100°C) we can compare the log porosities with experimental compaction of similar sands in the laboratory. There will then be a marked effect of overpressure reducing the effective stress. Poorly sorted sands will lose most of their porosity at relatively shallow depth but well sorted sand may have 30–35% porosity at 2–3 km depth, which corresponds to experimental compaction at about 20–30 MPa effective stress. This suggests that there is little creep over long geologic time.

Data from deep wells will always show a trend towards lower porosities with depth but there may also be intervals where this trend is reversed. This does not mean that net porosity has been created by diagenetic processes. Because of the very low solubility of silica and even more so of aluminium in porewater it is very difficult to explain how minerals in several metre thick sandstones can be dissolved without precipitation of other minerals in the same sandstones. Each lithology will have a characteristic porosity depth trend (Fig. 4.16) and increases in porosity reflect variations in the primary composition.

As discussed above the rate of quartz cementation can be modelled if the surface area available for quartz cementation and the time-temperature history during burial are known.

At about 4.0 km burial depth (120–140°C) the amount of quartz cement may be 10–15% so the remaining porosity may be only 10–15%. We do however find good reservoir quality (>20% porosity) at greater depths and higher temperatures but this is due to grain coatings. Prediction of porosity at great depth therefore requires that the occurrence of coating of chlorite, illite, haematite or microquartz can be predicted. Such prediction of the primary sediment composition must again be linked to facies and provenance.

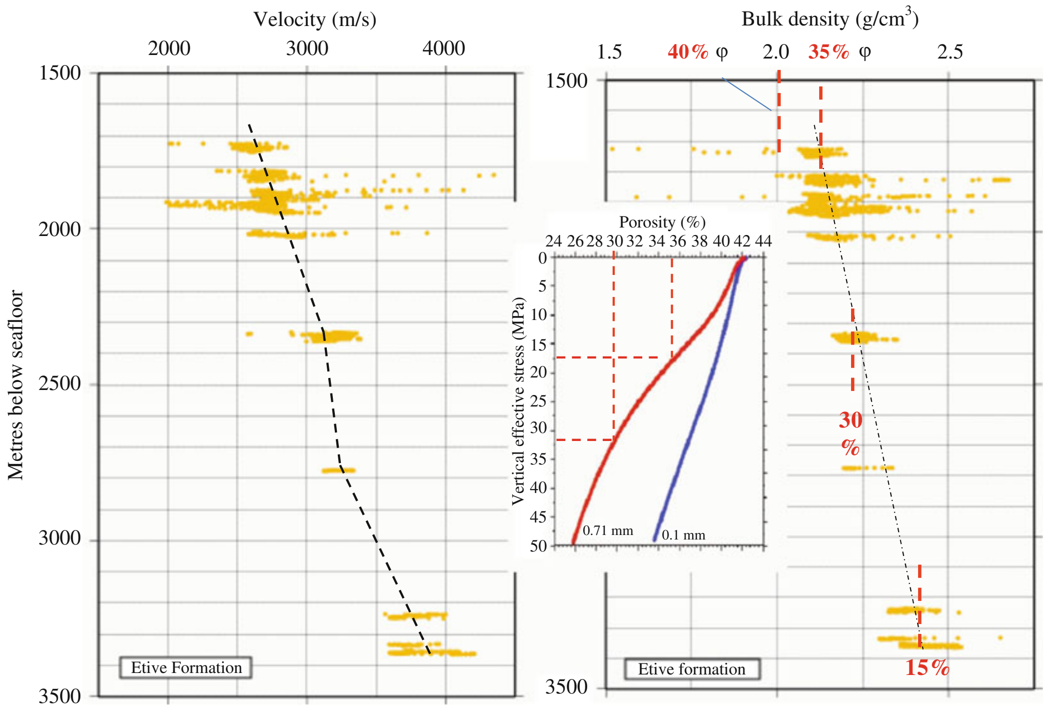

Fig. 4.16 Velocity/depth and density/depth trends for the Etive Fm (Brent Group) showing that a single lithology has a nearly linear trend with depth, based on Marcussen et al. (2009). The calculated porosities show that the compaction down to about 2 km depth is mechanical and similar to experimental data inserted from Chuhan et al. (2002). At greater depth, compaction is mostly chemical and higher compared with mechanical compaction

Prospects at great depth called HTHP (High Temperature, High Pressure) are expensive to drill and involve high risks, but in many mature basins nearly all the shallower prospects have been drilled already.

An extensive study of reservoir sandstones in the Gulf of Mexico showed that temperature and time are the main factors controlling reservoir quality (Nadeau et al. 2008).

4.17 Practical Prediction of Reservoir Quality

The most important factor controlling reservoir quality at depth is the primary composition of the sandstones. Sedimentological and sequence stratigraphic analyses are normally used primarily to predict reservoir geometry, but for diagenetic processes grain size, sorting and mineralogical composition are more critical. It is imperative to establish changes in provenance since the primary mineralogy places important constraints on diagenetic reactions at depth.

Reconstructions of facies and climate will provide a basis for predicting the degree of feldspar dissolution and precipitation of pore-filling kaolinite. Biogenic components like calcareous and siliceous organisms will control the distribution of carbonate cements and opal A, which will be altered to opal CT and quartz. Primary aragonite may cause extensive calcite cementation that occludes much of the primary porosity. Organic silica, e.g. from siliceous sponges, may serve as a precursor to grain-coating microquartz preserving porosity at depth.

Porosity loss due to mechanical compaction can vary greatly as a function of textural and mineralogical composition. Experimental compaction of loose sands with different grain size and sorting provides a good basis for prediction of porosity and inter-granular volume before quartz cementation. In cold sedimentary basins (low geothermal gradient) sand may be buried to 4–5 km before there is significant quartz cementation and in the absence of overpressure it can be subjected to 40–50 MPa effective stress.

Well log data from distinct lithologies buried to different depths may also provide a useful database for predicting the porosity loss due to mechanical compaction.

4.18 Burial Diagenesis and Reservoir Quality

Porosity loss due to quartz cementation can be modelled as a function of temperature, time and surface area available for quartz cementation. This is sensitive to grain size and grain coatings, which must be predicted from primary facies evaluation. The source of silica for quartz cementation may be unstable silica minerals like opal A, opal CT, grain contact dissolution or clay mineral reactions derived locally. The presence of pore-filling illite depends on precursors which may be smectite or kaolinite. Dissolution of kaolinite and precipitation of illite requires temperatures above 130°C and the local presence of K-feldspar. Sandstone containing mostly plagioclase does not develop pore-filling illite as there is insufficient supply of potassium. Prediction of reservoir quality can thus be based on provenance.

The examples of reservoir quality predictions listed above are based on the assumption that burial diagenetic reactions are essentially isochemical. Open system diagenesis allowing large scale import and export of solids in solution violates constraints from mineral solubilities and fluid flow rates and therefore provides a poor basis for prediction.

Statistical analyses show that the porosity is lost as a function of depth and that below 4.0-4.5 km burial depth the reservoir quality is in most cases insufficient for economic production. There are however exceptions where relative good reservoir quality is preserved despite burial to great depth (5-6 km). This may be due to low geothermal gradients and high burial rates. Clay coating is also an important factor retarding quartz cementation and porosity loss. High temperatures will also increase the degree of cementation in the shales and the permeabilities of seals may become so low that the pore pressure may reach fracture pressure (Nadeau et al. 2005). See Fig. 4.17.

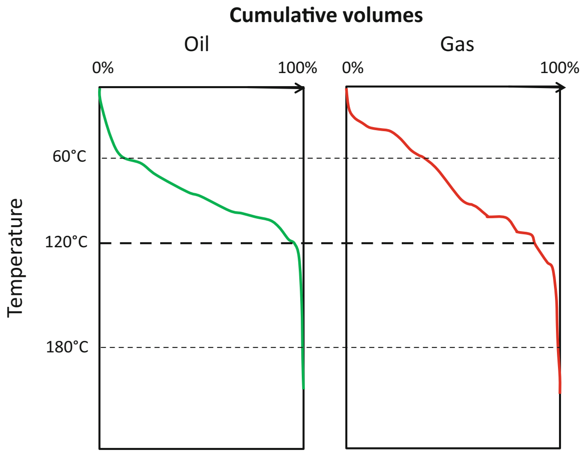

Fig. 4.17 Global cumulative percent reserves as a function of present reservoir temperature for (a) oil, (b), gas,. From Nadeau et al. 2005. Approximately 85% of the oil reserves occur between 60 and 120°C, and significant volumes of gas are found shallower than these temperatures. Many of these resevoirs have however been buried more deeply to higher temperatures.

4.19 Relationships Between Depositional Environment, Diagenesis and Rock Properties

Changes in the physical properties of sandstones from deposition to deep burial and uplift depends on the primary sediment composition with respect to textural and mineralogical composition. It is therefore important to study diagenesis as an integrated part of basin analysis. The source (provenance) of the sediments determines the starting composition with respect to the size of quartz grains from weathering and erosion of both igneous rocks and from older sedimentary rocks.

The feldspar content, and the composition of feldspars, control to a large extent diagenetic reactions in sandstones at greater depth. We have seen above that sandstones rich in plagioclase will react very differently from those rich in K-feldspar. Changes in mineral composition are also very important to reconstruct drainage and transport in relation to facies in sedimentary basins. It is unfortunate that many sedimentological descriptions include very little about the mineralogical and textural composition of sedimentary sequences from outcrops or cores. This information is critical for the prediction of reservoir quality at greater depth.

Prediction of mechanical compaction of sediments must be based on experimental compaction of sediments with different mineralogical and textural composition. Grain size, shape and sorting both in sandstones and mudstones play an important role in determining the mechanical compaction prior to chemical compaction. Interpretations of provenance and depositional environment should be used as input for prediction of mechanical compaction based on experimental compaction curves, thus linking diagenesis to more general basin analysis.

At shallow depth sediments may change their bulk composition due to meteoric water flushing causing leaching (dissolution) or evaporation (precipitation). Below the reach of significant meteoric water flow the porewater flow is very restricted with respect to transport of solids in solution.

Prediction of changes in rock properties due to chemical compaction is based on thermodynamics and kinetics and the initial mineral composition is then critical. Fluid flow and transport of solids in sedimentary basins can be constrained by modelling sediment compaction and also by observations of compositional stratification of the porewater. Burial diagenetic reactions are nearly isochemical, and mineral reactions can therefore be written as balanced equations. The burial diagenetic reactions are then functions of the primary sediment composition which can then be related to depositional environments and provenance.

Significant increased porosity (secondary porosity) is dependent on the dissolution and removal of solids in solution which may occur during freshwater flushing. Below the reach of freshwater the porewater flow is limited and a geochemically closed system cannot produce net increases in secondary porosity.

Prediction of rock properties such as porosity and velocity at a certain depth in a sedimentary basin must be based on the burial history (effective stress and temperature), but the primary mineralogical and textural composition of the sediments is equally important. Sedimentological studies of depositional environments and provenance should therefore be integrated with diagenesis and be a part of basin analysis which is used for basin modelling (Bjørlykke 2014).

4.20 How Much Oil Can Be Produced from Sandstone Reservoirs?

Petroleum exploration requires predictions about the reservoir properties ahead of drilling to justify the investment that a well represents. The reservoir properties determine the percentage of recoverable petroleum in each volume of rock. Even after drilling several exploration wells and also production wells in the development of an oil or gas field, the porosities and permeabilities are only known from the cores. Data from cores, cuttings and well logs must be extrapolated to produce a 3D model of the large volumes of rock between the wells. This must be based on predictions from facies distribution, distribution of faults and fault properties. Changes in reservoir properties as a function of depth require diagenetic models which can predict changes in porosity as a function of effective stress and temperature/time.

To calculate the producible oil in a prospect the total rock volume (Gross – G) in the structure must be estimated as well as the percentage of sandstone which can be produced (Net Sand – N).

The oil in place (V_p) is:

$$V_r \cdot \mathrm{N/G} \cdot \varphi\, O_{\mathrm{sat}}$$

Here V_r is the volume of oil between the oil/water contact (OWC) and the reservoir cap rock or the gas/oil contact. N/G (net/gross) is the ratio between the fraction of the reservoir rock that can be produced and the total volume of the reservoir rock. φ is the average porosity of the producible part of the reservoir (net volume). O_{sat} is the average saturation of oil; typically about 80–85% of the pores in sandstone are filled with oil. The remaining portion is water, which in siliciclastic rocks occupies the mineral surfaces and the smallest pores where the capillary entry pressure is too high for oil.

If the porosity is low the permeability will in most cases also be very low so that the flow of oil from the rock formation to the well becomes too slow to be economical. About 10% porosity may be the minimum porosity for defining the producible (net) part of the reservoir. In fractured reservoirs the permeability may be high even when the porosity is below 10%.

In the planning of production from an oil field, data from cores and logs from wells must be extrapolated into a 3D model of the flow properties of the reservoir.

The producible percentage of the oil in place is called the *recovery factor*, which may range from 20–30% up to 40–60%. In sandstones with good reservoir quality, improved production technology has in some cases (e.g. Statfjord and Gullfaks fields, offshore Norway) boosted the recovery factor to close on 70%. Recovery is limited both by the amount of oil remaining in the pores of the drained sandstones and the presence of undrained compartments within the reservoir where oil is bypassed. Reservoir quality is a very important factor in the financial risk assessment calculations for a prospect.

4.21 Conclusions

Diagenetic reactions are driven towards higher mechanical and chemical stability. Reactions in sandstones are driven by the effective stress from the overburden causing reduced porosity (volume) at temperatures below 70–80°C. At greater depth (higher temperatures) compaction is mostly chemical and mineral reactions are controlled by thermodynamics and kinetics. Because of low kinetic reaction rates (high activation energies) silicate reactions are very sensitive to temperature. The precipitation of quartz is slow and has high activation energy, and temperature is the main control on quartz cementation causing much of the porosity loss in well-sorted sandstones. The dissolution of K-feldspar and kaolinite at about 130°C occur because the mineral assemblage illite and quartz is more stable.

Because of the low solubility of silicate minerals and the limited flow of porewater in the deeper parts of sedimentary basins, burial diagenetic reactions must be nearly isochemical. Significant amounts of solids can not be exported from a sandstone and the porosity of a single reservoir rock will only decrease and can not increase during progressive burial.

Prediction of reservoir quality at great burial depth depends on the initial sediment composition (provenance), sedimentary facies (Fig. 4.18), early diagenetic processes and the subsequent burial history.

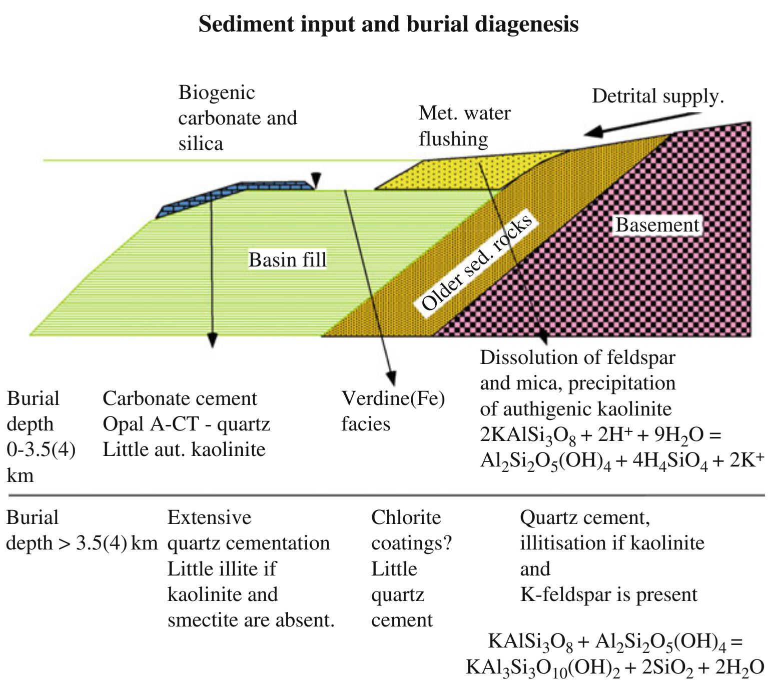

Fig. 4.18 Summary diagram for clastic diagenesis. The primary composition of the sand depends on the provenance, weathering, transport and depositional environment. Early diagenesis including meteoric flushing or marine cementation, is controlled by the depositional environment. Burial diagenesis is controlled by mineral stability (thermodynamics) and reaction rates (kinetics)

This should therefore be linked to facies models and provenance in addition to basin subsidence and heat flow. Modelling of quartz cementation has proved to be very useful for the prediction of porosity as a function of the temperature history of the reservoir sandstones, but the results are sensitive to the primary textural and mineralogical composition of the sandstones and the presence of grain coatings.

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