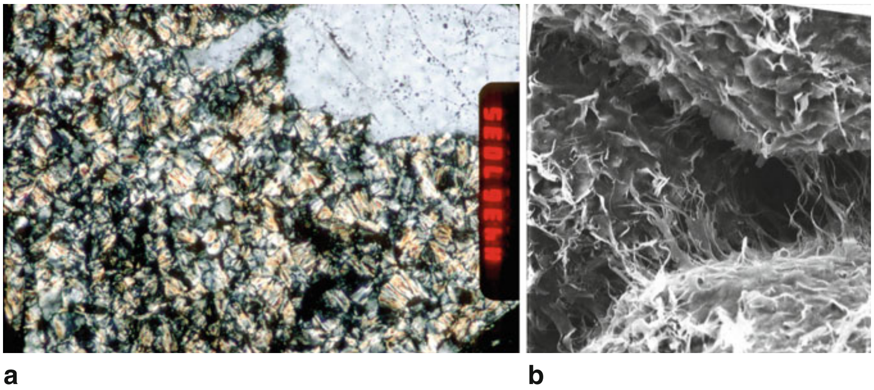

Fig. 4.14 (**a**) Pore-filling illite replacing pore-filling kaolinite preserving the kaolinite textures in Jurassic sandstone, Haltenbanken, offshore mid-Norway. (**b**) Pore-filling illite probably altered from smectite. Triassic sandstone from the North Sea

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.

$$\begin{array}{llcll} KAlSi_3O_8 + & Al_2Si_205(OH)_4 & = & KAl_3Si_3O_{10}(OH)_2 + & 2SiO_2 + H_2O \\ \text{K-Feldspar} + & \text{Kaolinite} & = & \text{Illite} \quad + & \text{Quartz} \end{array}$$

This is a simplified presentation of this reaction. 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 1983).

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 (3×10^{-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 (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 Prediction of Reservoir Quality

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.

4.15 Turbidites

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 Palaeocene 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.