

have been hanging to the pore wall and may have been pushed aside by growing quartz cement.

The isotopic composition ($\delta^{18}\text{O}$) 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 ($\text{Al}(\text{OH})_3$) 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 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. 2002, 2003). 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 (80–100°C) (Fig. 4.6). At greater depth, compaction is not primarily a function of effective stress but 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.

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.

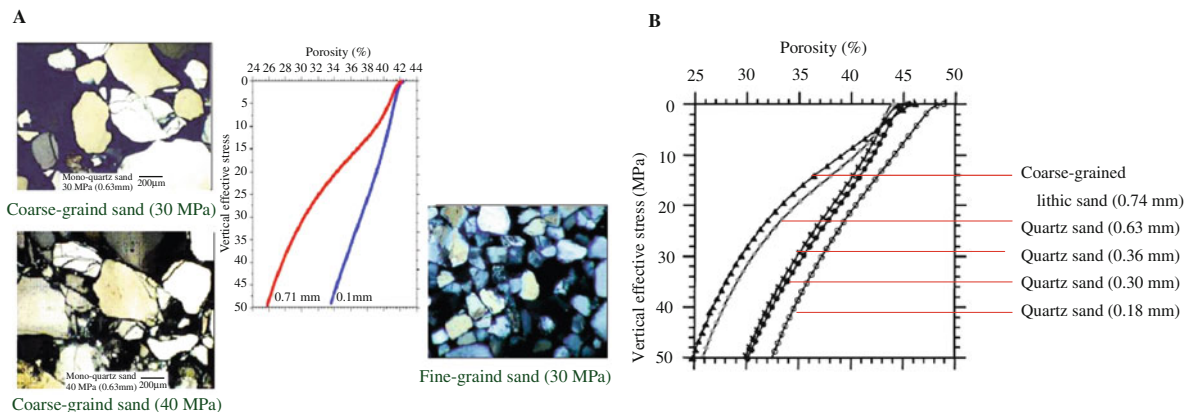


Fig. 4.5 (a) Experimental compaction of fine-grained and coarse-grained sand showing that well sorted fine-grained sand is less compressible than coarse-grained sand. (b) The porosity loss as a function of grain size due to more grain crushing (from Chuhan et al. 2007)

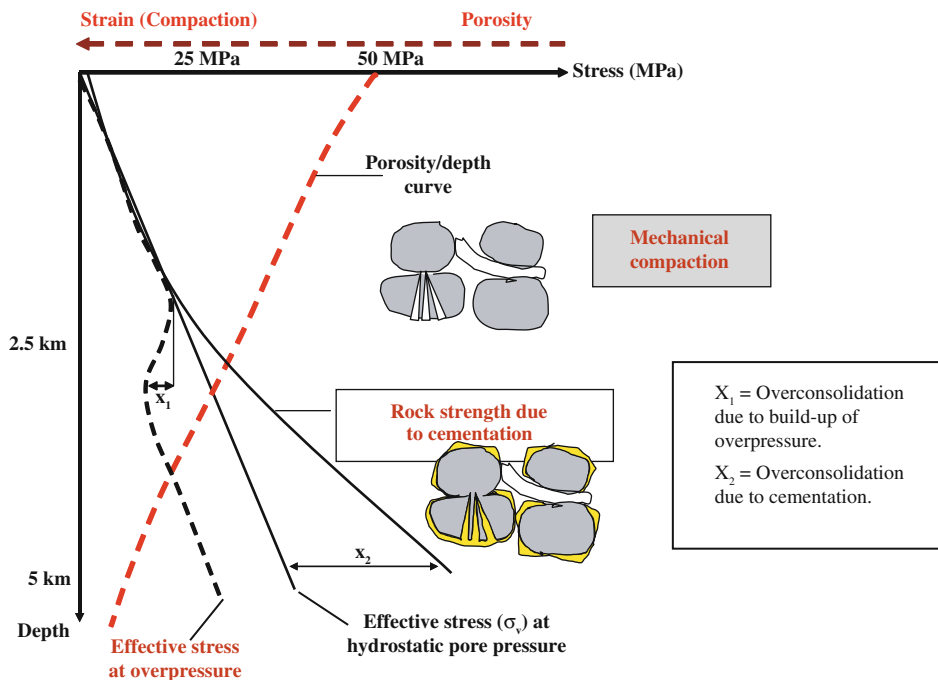


Fig. 4.6 Before sandstones become cemented (at 80–100°C) they compact mechanically as a function of effective stress (depth) by grain reorientation and grain breakage. Relatively small amounts of quartz cement (2–4%?) make the sandstone stiffer and “overconsolidated” so that there is little mechanical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

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. 4.5). 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.

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 further 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 micro-quartz 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 micro-quartz precipitated at low temperature (60–80°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 micro-quartz 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 micro-quartz.

At temperatures above about 100–120°C some of what we have called kaolinite has recrystallised 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 Ca²⁺ 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

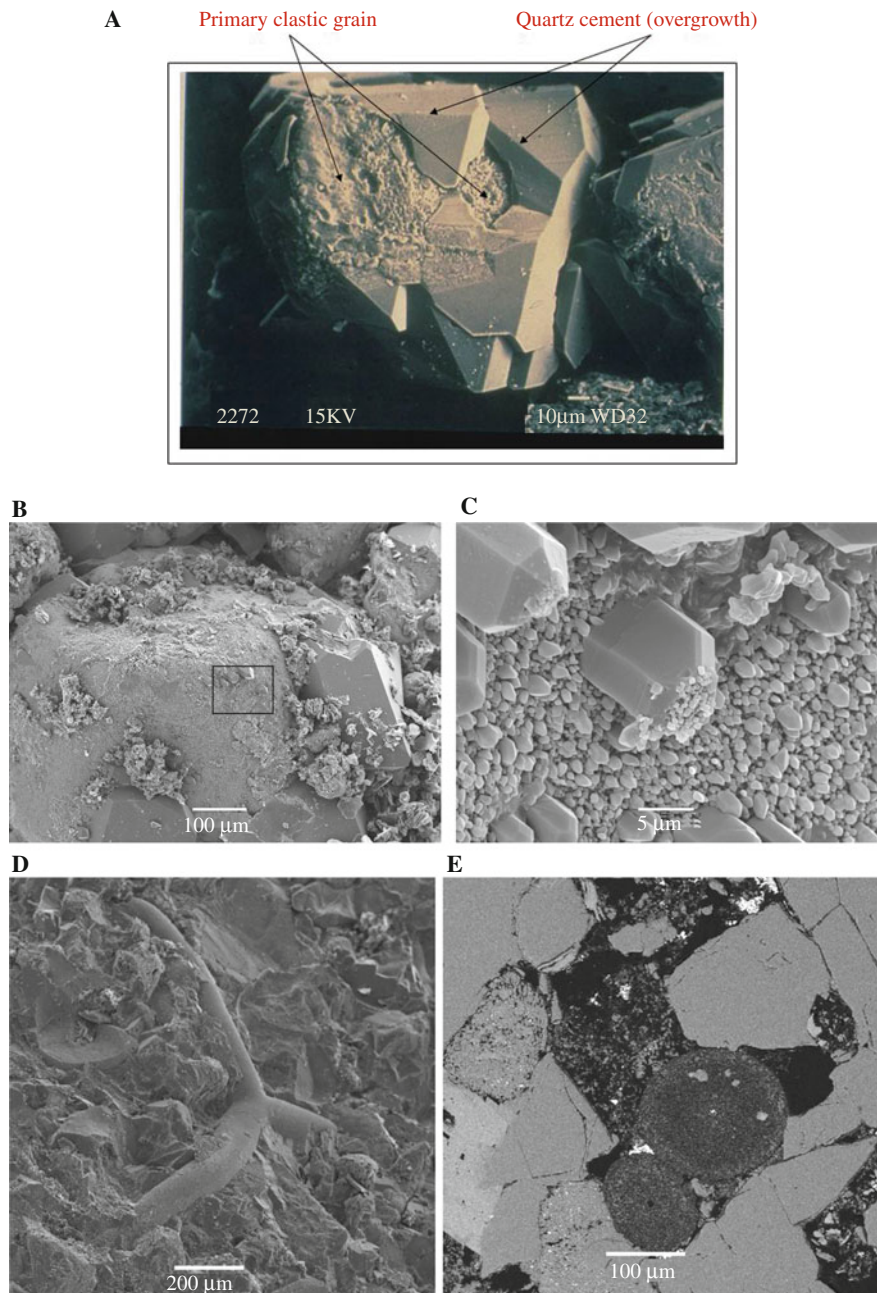


Fig. 4.7 (a) SEM image of quartz cement which has grown on detrital quartz grains into the pore space. Note that the authigenic quartz crystals have smooth crystal surfaces while the primary quartz grain has an irregular abraded surface from weathering and transport of the sand grain. Scale bar 10 µm (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

supersaturation caused by dissolution of organic silica (sponge spicules). (c) is an enlargement of Fig. 4.4b showing micro-quartz 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)