Chapter 4

Sandstones and Sandstone Reservoirs

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4.1 Introduction

About 60% 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.

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

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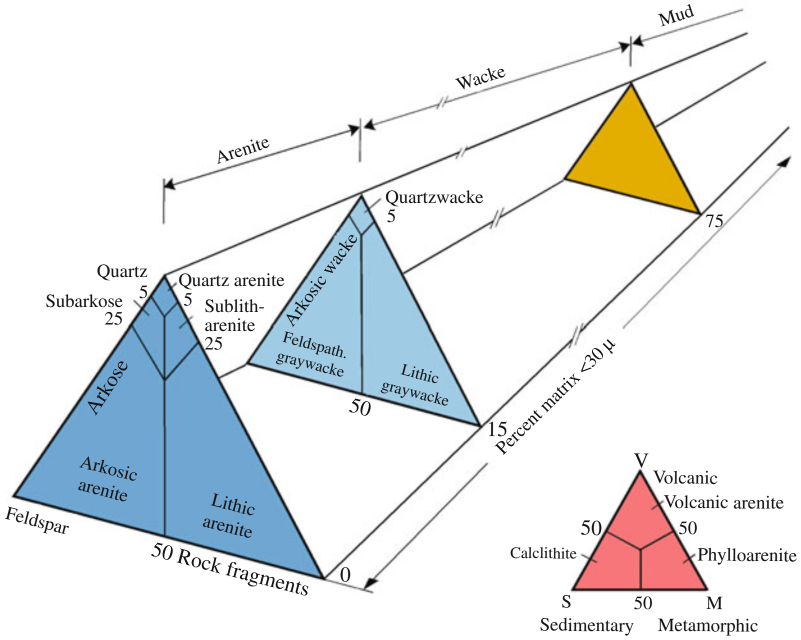

Fig. 4.1 Classification of sandstones (from Dott 1964)

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. 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 ole.

(4) Precipitation of cement (i.e. 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. 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.

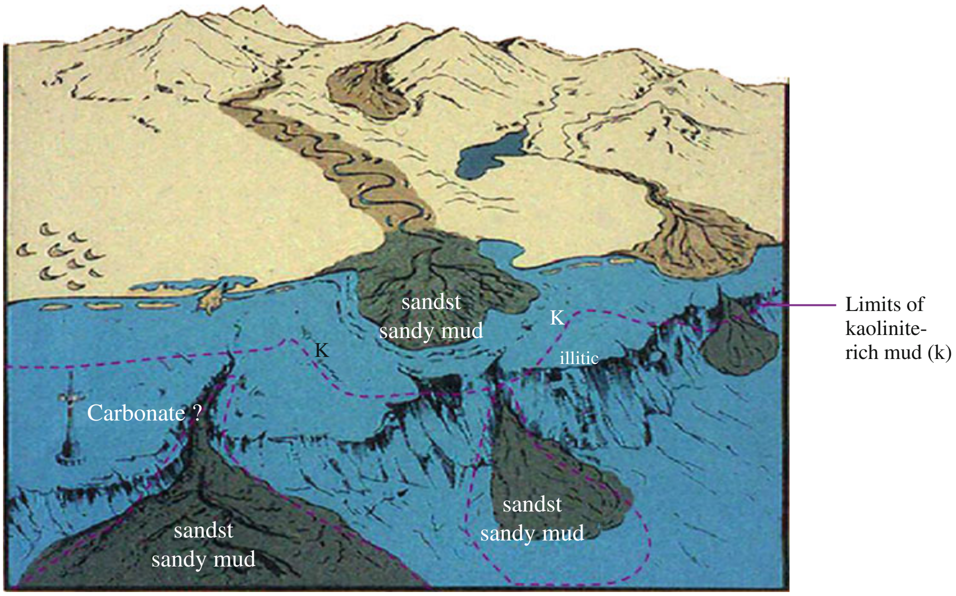

Fig. 4.2 Schematic illustration of a sedimentary basin on a continental margin. The primary composition of the sediments is a function of the provenance, transport and depositional environment. Fluvial, deltaic and shallow marine sediments will be flushed by meteoric water after deposition, particularly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

4.3 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