6.2.1 Introduction

The quantitative formulation of transport processes of energy (e.g., heat) or mass generally involves relating the driving force and the resulting flux with an appropriately chosen coefficient of proportionality:

\[
\text{Mass or energy flux} = \text{coefficient} \times \text{driving force}
\]

The most relevant transport processes in sedimentary systems are:
1. Heat flow (Fourier’s law)
2. Pressure-driven fluid flow (Darcy’s law)
3. Diffusive transport (Fick’s law)

The term “flux” denotes the quantity (of mass or energy) transported per unit time perpendicularly across a unit cross section area. In addition, coupled transport processes such as thermo-diffusion or thermal convection may occur. Theoretically, these coupled transport processes can be described by the Onsager reciprocal relations, used in non-equilibrium thermodynamics (cf. Mitchell 1976; Clennell 1997). Practically, however, the identification and quantitative description of coupled processes in the complex evolution of sedimentary systems is very difficult. Sedimentation and compaction processes, representing a combination of solid and fluid mass transfer in a gravitational field, constitute the major mass transport in sedimentary basins on the geological time scale (see Box 6.2.1). Therefore, their quantitative description and reconstruction is one fundamental aspect in numerical basin modelling.

Fluid transport in sedimentary systems may proceed either within the interconnected pore system (matrix transport) of sedimentary sequences (unconsolidated material or consolidated rocks), within vugular void spaces as in certain carbonate rocks, or along fault and fracture systems. The corresponding transport regimes differ substantially in terms of scale and transport efficiency. While pressure-driven fluid transport is the more efficient transport mechanism, diffusive transport of molecules and ions is of great relevance in the redistribution of matter over short distances (microns up to hundreds of metres on the geological time scale) and in (dead-) volumes that are not accessible to pressure-driven flow. Diffusion is frequently the rate-determining process in mineral and diagenetic reactions.

This chapter addresses fluid and heat transport processes in sedimentary systems and the current state of knowledge concerning their quantification. The focus is on transport in the pore and fracture systems, under the aspects of porous reservoir rocks and low-permeability seals.

6.2.2 Physical mechanisms and concepts

6.2.2.1 Overview

The ability of sedimentary rocks to accommodate and transport fluids is directly linked to their geological and tectonic history. Geological conditions such as depositional environment control the primary mineralogical composition, grain size, grain size distribution and grain shapes. The texture and pore system of rocks are subsequently modified by compaction, cementation and diagenetic or metamorphic overprint so that the petrophysical rock properties such as porosity, permeability and tortuosity are affected by numerous processes acting simultaneously or sequentially. Theoretical and practical aspects of the petrophysical characterisation of rocks are discussed in numerous textbooks (e.g., Bear 1972; Freeze and Cherry 1979; Tiab and Donaldson 2004) and therefore only a short recapitulation of the most important terms and concepts will be given here. These are:

- porosity ($\Phi$): a measure of the void space of porous media
- permeability ($k$): a measure of the ability of porous rocks to transport fluids
- tortuosity ($\tau$): a measure of the structure of interconnected pore channels and interstices.

The endogenic and exogenic mass transport processes, such as erosion and deposition are not considered in this list.
Box 6.2.1. Mechanical and physicochemical compaction

Compaction of sediments is the reduction in volume of a sedimentary body that is caused by stress. This stress commonly results from the increasing overburden load of deposits accumulated during subsidence of a sedimentary basin. Compressional stress owing to tectonic movements may also cause compaction. Compaction involves a combination of mechanical and physicochemical processes and results in the decrease of initial porosity, permeability, and stratigraphic thickness, and the increase in density.

Overburden stress is transmitted by grain-to-grain stress in uncremented sediments. The mechanical response to this stress includes the rearrangement/rotation of components, the plastic deformation of mechanically unstable particles, and the brittle failure of grains. Physicochemical compaction includes dissolution of components at points of contact and mineral transformations that result in a net reduction of bulk volume. The loss of porosity due to compaction is accompanied by expulsion of pore fluid. If the pore water pressure is equal to the hydrostatic pressure at any depth, and the sedimentary column is completely carried by the grain framework, the sediments are normally consolidated. If low permeability prevents the escape of pore fluid, compaction is retarded and the sediments are underconsolidated. Parts of the overburden load are then carried by the pore water, which builds up excess pore water pressure. Overconsolidation results from erosion of parts of the overburden, thus the sediments reflect a state of consolidation that is higher than expected from the remaining sedimentary column.

Mechanical compaction in siliciclastic sediments is generally most effective in the uppermost 1-1.5 km of burial (Fig. 1). Clay-rich sediments have higher initial porosities than sand-rich sediments, but the latter are much less affected by mechanical compaction (Fig. 1). Sandstones rich in rigid grains such as quartz become less strongly compacted than sandstones rich in mechanically unstable lithic grains such as (meta-)pelites or volcanic clasts. Physicochemical compaction in clays involves dehydration of clay minerals, notably the progressive transformation of smectite to illite. Physicochemical compaction in sandstones takes place by dissolution of components at grain-grain contacts (“pressure solution”). Pressure-related dissolution of quartz is most effective at burial depth greater than about 2 km, but is also controlled by temperature, mineralogy and rock fabric (e.g., Houseknecht 1988; Bjørkum 1996). Pressure solution of carbonate is important for compaction of carbonate-rich sands and limestones.

The interaction of mechanical compaction, physicochemical compaction and cementation is difficult to predict and leads to a wide scatter in measured porosity-depth relationships for individual case studies (e.g., Rieke and Chilingarian 1974, Chilingarian and Wolf 1975). To approximate compaction in sediments, Athy (1930) proposed the negative exponential equation:

$$\Phi_z = \Phi_0 \cdot e^{-cz} \quad (6.2.1)$$

where $\Phi_z$ is the porosity at the depth $z$, $\Phi_0$ is the initial porosity, $z$ is the burial depth, and $c$ is an empirical coefficient depending on the type of sediment. As a result of compaction, sandstones typically have higher porosities than shales at elevated burial depths. However, the reliability of porosity-depth curves is very limited if diagenetic processes and the heterogeneities in terms of mineralogical composition, grain size and sorting are not adequately considered. For shales, refined models have been developed to better account for clay mineral dehydration and permeability evolution (e.g., Broichhausen et al. 2005). Sandstone porosity may be strongly affected by cementation and/or dissolution. Early cementation, for example, strongly reduces porosity, but effectively hampers further compaction. Sandstone compaction can be more precisely described by intergranular volume, i.e., porosity plus volume of cements and matrix present in between the framework grains (Fig. 1). Intergranular volume vs. depth curves can help to predict compaction and porosity evolution for sandstones (Landar and Walderhaug 1999; Paxton et al. 2002).

![Figure 1. Schematic intergranular volume-depth relationships of sandstones and porosity-depth relationships of shales caused by mechanical and physicochemical compaction during burial.](image-url)

a) Example for expected percentage of mechanical compaction of a well sorted, medium-grained, quartz-rich sandstone (A); reduction of intergranular volume by physicochemical compaction (B); resulting intergranular volume-depth relationship (C); grey envelope represents the range of intergranular volume-depth relationships of various sandstones. b) Example for expected percentage of mechanical compaction of a shale (A); reduction of porosity by dehydration of clay minerals (B); resulting porosity-depth relationship (C); grey envelope represents typical porosity-depth relationships of shales. In nature, clay compaction is probably rather an episodic than a continuous process (e.g., Cartwright 1994)
6.2.2.2 Porosity

Porosity is the ratio of the pore volume of a rock $V_p$ to its bulk volume $V_b$:

$$\Phi = \frac{V_p}{V_b}$$  \hspace{1cm} (6.2.2)

$$\Phi = \frac{V_g - V_{gr}}{V_g}$$  \hspace{1cm} (6.2.3)

where $V_{gr}$ is the grain volume. Porosity is frequently reported as a percentage rather than as a volume fraction. Unconsolidated clay-rich (pelitic) sediments may have porosities up to 70% shortly after deposition while porosity values of gravels and sands are usually below 50%. Upon compaction and consolidation during burial and diagenesis the porosity is successively reduced and may approach values close to zero in deep shales and crystalline rocks. Typical porosity ranges for unconsolidated sediments and rocks are listed in table 6.2.1.

During compaction the pore volume of sediments may be altered due to fluid-solid interactions (see Schöner et al. this volume): primary porosity refers to the porosity retained during burial and affected only by physical compaction; secondary porosity is additional porosity created by dissolution processes or fracture formation. Due to sedimentation and diagenetic processes, initially open and interconnected pores may become isolated. Thus, two further categories, the total (absolute) and the effective porosity have to be discriminated: the absolute porosity is the ratio of the total void space in the sample to the bulk volume, regardless of the degree of interconnection of the pores. It encompasses the volume of fluid-filled isolated pores, adsorbed water on grains or particle surfaces, e.g., clay. Effective (accessible) porosity thus denotes the fraction of the pore system that is interconnected and amenable to fluid exchange. This porosity may be further subdivided into “transport porosity”, the network of interconnected pores through which the volume flow of fluid phases occurs, and “dead-end” pores with stagnant fluid where exchange is possible only by diffusion (Fig. 6.2.1).

Factors affecting permeability

The ability of porous rocks to conduct fluids depends on porosity and the degree of interconnection of the pores. Layering and rock texture are controlled by depositional and diagenetic conditions and usually result in a smaller

Table 6.2.1. Range of porosity values for unconsolidated deposits and rocks. Porosity is expressed as fraction

<table>
<thead>
<tr>
<th>Unconsolidated deposits</th>
<th>Rocks</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gravel</td>
<td>0.4 - 0.25</td>
</tr>
<tr>
<td>Sand</td>
<td>0.5 - 0.25</td>
</tr>
<tr>
<td>Silt</td>
<td>0.5 - 0.25</td>
</tr>
<tr>
<td>Clay</td>
<td>0.7 - 0.4</td>
</tr>
<tr>
<td>Fractured basalt</td>
<td>0.5 - 0.05</td>
</tr>
<tr>
<td>Karst limestone</td>
<td>0.5 - 0.05</td>
</tr>
<tr>
<td>Sandstone</td>
<td>0.3 - 0.05</td>
</tr>
<tr>
<td>Dolomite</td>
<td>0.2 - 0.0</td>
</tr>
<tr>
<td>Shale</td>
<td>0.1 - 0.0</td>
</tr>
<tr>
<td>Crystalline rocks</td>
<td>0.1 - 0.0</td>
</tr>
</tbody>
</table>

6.2.2.3 Permeability

If an interconnected pore or fracture system exists, fluids can pass through porous rocks under the influence of a pressure gradient. This property is denoted as permeability or, in hydrogeology, hydraulic conductivity. Permeability is a characteristic property of the porous medium and is determined by relating the volumetric fluid flow rate $q$ (m³/s) through a given cross section area $A_c$ (m²) to the pressure gradient, $dp/dx$ (Pa/m) acting on the fluid. This relationship, commonly known as Darcy’s law, is written as:

$$v = \frac{q}{A_c} = \frac{k}{\mu} \left( \frac{dp}{dx} \right)$$  \hspace{1cm} (6.2.4)

where $v$ (m/s) is the volume flux or “Darcy velocity”, $\mu$ (Pa s) the viscosity of the fluid and $k$ (m²) is the permeability coefficient. Quite frequently permeability coefficients are reported in the unit “Darcy” rather than in the SI unit m². One Darcy (1D) is equal to 0.9869 µm² (0.9869·10⁻¹² m²). Because permeability coefficients of sedimentary rocks extend over a range of more than 10 orders of magnitude, the terms millidarcy mD (~10⁻¹⁵ m²), microdarcy µD (~10⁻¹⁸ m²) or nanodarcy nD (~10⁻²¹ m²) are used. Table 6.2.2 lists the ranges of permeability coefficients of unconsolidated sediments and sedimentary rocks.

The above definition of permeability (absolute permeability) assumes complete saturation of the pore system with one single fluid. If more than one fluid phase is present in the pore system, e.g., in oil/water and gas/water systems, the transport rates of the individual phases depend on their relative volume fractions (saturation) in the pore systems and are described in terms of relative permeabilities. The sum of the individual phase permeability coefficients, the effective permeability coefficient, is always less than the (single phase) absolute permeability coefficient of the system.

Factors affecting permeability

The ability of porous rocks to conduct fluids depends on porosity and the degree of interconnection of the pores. Layering and rock texture are controlled by depositional and diagenetic conditions and usually result in a smaller
or larger degree of permeability anisotropy (Fig. 6.2.2). Furthermore, permeability exhibits significant stress dependence as a result of pore volume decrease and fracture closing when rocks are exposed to high effective stress. The interaction of aqueous fluids with clay (swelling of smectites) may considerably reduce conduits or may even block transportation. Thus, permeability also depends on the chemical composition of the flow medium (brine). Therefore, laboratory based permeability measurements must consider the environmental conditions of the samples to provide reliable results of the in-situ permeability.

6.2.2.4 Permeability-porosity relationships

Because permeability depends on the connectivity of the pore system rather than on absolute porosity, a general relationship between porosity and permeability does not exist. However, semi-empirical and empirical correlations can often be established for individual formations or lithologic units. In a semi-logarithmic plot of permeability versus porosity one often finds “straight” lines with different slopes for different rock types, e.g., coarse-grained to clay-rich sandstones or various carbonates. These differences in slopes reflect differences in grain size and grain distribution, fabric elements, texture and cementation.

More sophisticated models take into consideration additional parameters such as grain-size distribution, specific pore surface area and pore dimensions (Table 6.2.3). Grain-based models correlate permeability with the square of grain size. Grain sorting may be used as an additional parameter. Surface area models indicate that permeability is proportional to the inverse square of specific pore surface area, and pore-dimension models show permeability proportional to the inverse square of a pore dimension.

The most popular and fundamental porosity-permeability correlation model is the Kozeny-Carman equation (Wyllie and Gregory 1955 and references cited therein). It expresses permeability as a function of two basic properties: porosity and specific surface area. In this model, the conduits are assumed to be capillary bundles. For a single cylindrical capillary of radius r the fluid flow rate is given by Poiseuille’s law:

\[ q = \frac{r^4 \pi}{8 \mu} \left( \frac{dp}{dx} \right) \left[ \frac{m^3}{s} \right] \]  (6.2.5)

The fluid flow rate \( q \) or flux \( v \) in a capillary bundle can be calculated by combining the Darcy (Eq. 6.2.4) and Poiseuille (Eq. 6.2.5) equations. One then obtains an expression of the following type:

\[ v = \frac{q}{\alpha L} = \frac{A}{\alpha \mu} \left( \frac{dp}{dx} \right) \]  (6.2.6)

Here \( A \) is the cross sectional area of a micro-channel, \( \alpha \) is a dimensionless constant taking into account the geometry of the pore channel. If the concept of hydraulic radius (= ratio of pore volume to solid-fluid interfacial area) is applied, then the cross sectional area \( A \) is replaced by tortuosity \( \tau \) \( \left( \tau = (L_e/L)^2 \right) \) and specific surface area \( S_{pore} \) of the pores, thus yielding an extended version of the Kozeny-Carman relation:

\[ k = \frac{\Phi^3}{K \tau S_{pore}^4 \left(1 - \Phi \right)^2} \]  (6.2.7)

**Figure 6.2.1.** Structures of porous rocks: a) well-sorted uniform granular material with microporous particles; b) poorly sorted graded granular material with small particles filling the large pores; c) partially closed discontinuity systems occurring in an intact porous rock; d) open discontinuity systems occurring in an intact rock due to mechanical fracturing. (after Freeze and Cherry 1979 and Meinzer 1923)
Here $K_T$ is the effective zoning factor, depending on pore size and shape, grain size and shape, their distribution, tortuosity and other factors. $K_T$ is a characteristic parameter for a given stratigraphic flow unit (see below).

Within one geological unit one may find flow conditions in the reservoir being laterally and vertically continuous, thus having similar permeability, porosity and rock fabric characteristics. Such a system can be assumed to be one flow unit. Due to sedimentation conditions different flow units may be found within a formation, e.g., as a function of clay content in sandstones. Different flow units can be hydraulically connected to other flow units. To map such units, field techniques as well as laboratory data can be used. Amaefule et al. (1993) proposed a method for characterising flow zones with similar hydraulic properties. The method uses a modified version of the Kozeny-Carman equation and the mean hydraulic radius concept. It is applied on scales from the pore level through the inter-well level and involves three terms that are convenient for use with conventional core analysis data:

- reservoir quality index $RQI = 0.0314 \cdot \frac{k}{\sqrt{\Phi}}$ (6.2.8)
- normalized porosity index $NPI = \frac{\Phi}{1-\Phi}$ (6.2.9)
- flow zone indicator $FZI = \frac{RQI}{NPI}$ (6.2.10)

where $RQI$ is given in µm. The $FZI$ includes geological attributes like texture and mineralogical composition as well as pore geometry. According to Tiab and Donaldson (2004), samples with the same $FZI$ values have similar pore throat size, and thus constitute a flow unit. Double-logarithmic plots of $RQI$ vs. $NPI$ result in straight lines exhibiting different slopes for different flow zones.

**Table 6.2.2.** Magnitude of permeability coefficients (in Darcy) of unconsolidated deposits and rocks

<table>
<thead>
<tr>
<th>Unconsolidated deposits</th>
<th>Fractured basalt</th>
<th>Karst limestone</th>
<th>Unconsolidated deposits</th>
<th>Fractured basalt</th>
<th>Karst limestone</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gravel</td>
<td>10$^{5}$</td>
<td>10$^3$-10</td>
<td>10$^{-4}$-10$^3$</td>
<td>10$^{-4}$-10</td>
<td>10$^{-4}$-10$^3$</td>
</tr>
<tr>
<td>Sand</td>
<td>10$^{-5}$</td>
<td>10$^3$-10</td>
<td>10$^{-4}$-10$^3$</td>
<td>10$^{-4}$-10</td>
<td>10$^{-4}$-10$^3$</td>
</tr>
<tr>
<td>Silt</td>
<td>10$^{-5}$</td>
<td>10$^3$-10</td>
<td>10$^{-4}$-10$^3$</td>
<td>10$^{-4}$-10</td>
<td>10$^{-4}$-10$^3$</td>
</tr>
<tr>
<td>Clay</td>
<td>10$^{-5}$</td>
<td>10$^3$-10</td>
<td>10$^{-4}$-10$^3$</td>
<td>10$^{-4}$-10</td>
<td>10$^{-4}$-10$^3$</td>
</tr>
<tr>
<td>Fractured basalt</td>
<td>10$^{-5}$</td>
<td>10$^3$-10</td>
<td>10$^{-4}$-10$^3$</td>
<td>10$^{-4}$-10</td>
<td>10$^{-4}$-10$^3$</td>
</tr>
<tr>
<td>Karst limestone</td>
<td>10$^{-5}$</td>
<td>10$^3$-10</td>
<td>10$^{-4}$-10$^3$</td>
<td>10$^{-4}$-10</td>
<td>10$^{-4}$-10$^3$</td>
</tr>
<tr>
<td>Sandstone</td>
<td>10$^{-5}$</td>
<td>10$^3$-10</td>
<td>10$^{-4}$-10$^3$</td>
<td>10$^{-4}$-10</td>
<td>10$^{-4}$-10$^3$</td>
</tr>
<tr>
<td>Dolomite</td>
<td>10$^{-5}$</td>
<td>10$^3$-10</td>
<td>10$^{-4}$-10$^3$</td>
<td>10$^{-4}$-10</td>
<td>10$^{-4}$-10$^3$</td>
</tr>
<tr>
<td>Shale</td>
<td>10$^{-5}$</td>
<td>10$^3$-10</td>
<td>10$^{-4}$-10$^3$</td>
<td>10$^{-4}$-10</td>
<td>10$^{-4}$-10$^3$</td>
</tr>
<tr>
<td>Crystalline rocks</td>
<td>10$^{-5}$</td>
<td>10$^3$-10</td>
<td>10$^{-4}$-10$^3$</td>
<td>10$^{-4}$-10</td>
<td>10$^{-4}$-10$^3$</td>
</tr>
</tbody>
</table>

Figure 6.2.2.
6.2.3 Fault seals and top seals

6.2.3.1 Overview

Top seals and fault seals have a major effect on the distribution of fluid pressure and accumulation of hydrocarbons in sedimentary basins, but many details of the processes and their dynamics are poorly understood. There is a reasonable understanding of the basic physical processes of retention, but the complex geometries and many feedback processes make quantitative prediction difficult.

A large majority of proven seals are evaporites, shales or organic rich rocks. They are laterally continuous, ductile and have high capillary entry pressures. Sealing performance is thought to depend on the ability to resist fracturing. A ranking by lithology has been proposed as Salt > Anhydrite > Organic-rich shales > Clay shales > Silty shales > Carbonate mudstones > Cherts (Downey 1984).

It is generally accepted that laterally continuous, ductile and impermeable rocks are favorable seals, and that lithology is the most important parameter in controlling seal quality. The capillary resistance of a few inches thick clay shale is sufficient to trap a large oil column. However, a thicker seal is beneficial because it provides more chance of lithological continuity and protects against faults which offset the cap rock by more than its own thickness.

For a full understanding of top seals and fault seals one has to consider both the macro- and microscale aspects of the system. At the macroscale these are the geometry of the structure, the lateral continuity of the sealing lithology and the complicated plumbing system in layers juxtaposed by faults (Fig. 6.2.3). At the microscale there are a number of mechanisms of single- or multiphase fluid flow through matrix- or fracture porosity (Fig. 6.2.4).

### Table 6.2.3. Permeability models

<table>
<thead>
<tr>
<th>Model</th>
<th>Author</th>
<th>Equation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sand pack</td>
<td>Krumbein and Monk (1943)</td>
<td>( k = 0.76 , D^2 , e^{-1.31} )</td>
</tr>
<tr>
<td>Grain-based</td>
<td>Berg (1970)</td>
<td>( k = 80.8 , D^2 , \Phi^{1.1} , e^{-1.30} )</td>
</tr>
<tr>
<td>Grain-based</td>
<td>Van Baaren (1970)</td>
<td>( k = 10 , D^2 , \Phi^{1.4} , C^{3.64} )</td>
</tr>
<tr>
<td>Surface area</td>
<td>Timur (1968)</td>
<td>( k = 0.136 , S_{wi}^{-2} , \Phi^{1.4} )</td>
</tr>
<tr>
<td>Surface area</td>
<td>Sen et al. (1990)</td>
<td>( k = 0.794 , T_{lo}^{1.18} , \Phi^{m+2.15} )</td>
</tr>
<tr>
<td>Pore size model</td>
<td>Kozeny-Carman in Carman (1956)</td>
<td>( k = 400 , R_c^2 , \Phi^{m} )</td>
</tr>
<tr>
<td>Pore size model</td>
<td>Winland in Pittman (1992)</td>
<td>( k = 49.4 , R_c^2 , \Phi^{1.47} )</td>
</tr>
<tr>
<td>Pore size model</td>
<td>Katz and Thompson (1986)</td>
<td>( k = 17.9 , R_c^2 , \Phi^{m} )</td>
</tr>
</tbody>
</table>

6.2.3.2 Micro- and macroscale processes

On the microscale, one can distinguish two hydrocarbon transport mechanisms through a seal: separate phase flow and diffusional transport (Fig. 6.2.4). Separate phase flow can occur in (i) relatively permeable seal lithologies, with
intermediate capillary displacement pressures, or in (ii) tight but brittle lithologies with connected fracture systems.

Capillary effects

In a water-wet porous rock the movement of oil or gas is hindered by the requirement of a pressure difference to force the hydrocarbon-water interface through the pore throats in the rock. For a rock with a given capillary entry pressure the maximum hydrocarbon column held by a capillary seal can be calculated if the properties of the hydrocarbon-water system are known. Watts (1987) describes applications of this principle to many field situations.

Darcy flow

When the top seal is not water-wet, or when its capillary entry pressure has been reached, the seal will leak hydrocarbons by flow through its matrix porosity. At sufficiently low permeabilities considerable hydrocarbon accumulations may be supported. The flux out of the trap is a function of hydrocarbon column length and in dynamic systems with simultaneous charge such a system is to some extent self-regulating. Important parameters here are the seal’s second phase saturation and relative permeability.

Flow through fractures

By their capillary resistance, tight shales can hold very large hydrocarbon columns. It is clear that hydrocarbon leakage can only occur by other mechanisms. Migration through fractures is a likely candidate. Here, the two main classes are shear and extensional fracturing.

Diffusion through the pore water

The solubility of methane and light hydrocarbons in water is of such magnitude that diffusion of gas through top seals may become an important mechanism of leakage through geologic time. In basins where generation and expulsion were active over a long geologic period, the pore water may become saturated everywhere. Under these conditions diffusional fluxes are controlled by the equilibrium concentration gradient in a hydrostatic column.

6.2.3.3 Application to fluid-rock systems

In a fluid-filled porous rock, capillary displacement of a second, non-miscible phase cannot be adequately described using a single pore throat diameter. This is done instead by specifying the capillary pressure curve which relates capillary pressure to saturation. The capillary displacement pressure \( P_d \) is defined as the fluid pressure which produces a significant saturation (about 10%) and corresponding second phase permeability. This concept is illustrated in figure 6.2.5.

Analogous to the arguments given above, the maximum hydrocarbon column held by a capillary seal can be calculated when the relevant rock parameters and fluid parameters are known. This has been done in the past by a number of authors (Berg 1975; Schowalter 1979; Watts 1987; Ingram et al. 1997).

In this case, the maximum hydrocarbon column \( h_{\text{max}} \) held by a specific seal at a given depth is given by:

\[
h_{\text{max}} = \frac{2P_{\text{gwc}}}{g \Delta \rho}\tag{6.2.13}
\]
where \( p_{\text{dis}} \) is the capillary displacement pressure for a hydrocarbon-water system.

Two points are worth noting here. Firstly, \( h_{\text{max}} \) to a first approximation is independent of overpressure in the water phase. Figure 6.2.6 shows pressure-depth plots around two seals assumed to have identical \( p_{\text{dis}} \) at different depths. Both seals hold the same hydrocarbon column by capillary resistance, but in the shallower one the pressure at top reservoir is very close to the minimum principle stress and the risk of hydrofracturing is high. Secondly, this analysis assumes full pressure communication between the wetting pore fluid in the lower part of the seal and in the reservoir. When the seal is slightly overpressured with respect to the reservoir, a larger hydrocarbon column can be held by the same seal. More generally, hydrodynamic trapping (Hubbert 1953) in basins with regional groundwater flow patterns can alter the trapping conditions and shapes of hydrocarbon-water contacts significantly.
Displacement pressure

Rock properties enter the calculations via the in-situ hydrocarbon-water displacement pressure $p_{dHC}$. This parameter is usually determined by conversion of data from high pressure mercury injection in air-dried samples taken from cores or larger drill cuttings (Fig. 6.2.7).

Direct measurement using gas and water in samples under in-situ stress and temperature conditions is possible but much more time-consuming than in mercury-air tests (Katz and Lee 1990). From the displacement pressure of the mercury injection test ($p_{dHg}$) the hydrocarbon-water displacement pressure is calculated as

$$p_{dHC} = p_{dHg} \left( \frac{\gamma_{HC} \cos \theta_{HC}}{\gamma_{Hg} \cos \theta_{Hg}} \right)$$

(6.2.14)

Figure 6.2.5. Schematic diagram illustrating the principle of capillary entry and associated second phase permeability

Figure 6.2.6. An illustration of the effect of different pore pressure systems on retention. PVT properties of the hydrocarbons are assumed to be independent of depth, and the displacement pressure of both seals is the same. Structural closure is larger than $h_{max}$. (a): the two seals hold the same hydrocarbon column, (b): the lower seal holds $h_{max}$ despite the overpressure, but the higher seal is breached by hydrofracturing. (c): the lower seal holds a longer hydrocarbon column then in (a) because of the downward flow out of the seal. The upper seal is inactive in this case due to lack of charge.
Capillary pressures and permeability in permeable seals may be coupled using the Leverett J function

\[ J(S_w) = \frac{p_c}{G(k/\Phi)^{0.5}} \]  \hspace{1cm} (6.2.15).

Two-phase flow can then be simulated using numerical techniques developed in reservoir engineering.

To get an impression of the range of possible values in \( p_{dHg} \), one can compare values from reservoir sandstones (\( p_{dHg} < 0.1 \text{ MPa} \)) to shales. As shown by Smith (1966), significant columns can be retained by rocks with \( p_{dHg} > 1 \text{ MPa} \). On the other end of the scale, in tight shales \( p_{dHg} \) can be in excess of 50 MPa: these rocks could hold such large columns that seal capacity is rarely reached and seal leakage, when present, is by other mechanisms (Fig. 6.2.8).

**Large-scale two phase flow**

Two phase flow at large scale is a very complex process. Figure 6.2.9 shows schematic illustrations of the interplay between capillary and geometric processes which can result in separation of the flow paths of oil and gas in the subsurface.

**Rock properties**

One of the major sources of uncertainty in capillary seal prediction lies in estimating the Mercury-air displacement pressure \( p_{dHg} \). A capillary seals database was presented by Hildenbrand et al. (2006) (Fig. 6.2.10), who derived an equation to calculate \( p_{dHg} \) from data on porosity and clay content, allowing models of depth-dependent seal capacity in basins. It should be noted however, that although figure 6.2.11 shows a reasonable correlation for the set of Tertiary clays studied, extrapolation to other clays is not possible. Figure 6.2.11 through 6.2.13 show examples of mercury-air data from the Buntsandstein and Early Tertiary of the Central European Basin Systems (CEBS).

**Gas density**

Gas density \( \rho_g \) at a given pressure and temperature may be approximated using the equation of state for a methane-ethane (\( \text{C}_1 - \text{C}_2 \)) mixture:

\[ \rho_g = \frac{M_g \gamma}{Z RT} \]  \hspace{1cm} (6.1.16)

where \( \rho_g \) is gas pressure, \( R \) is the gas constant (8.3144 J mol\(^{-1}\)K\(^{-1}\)), \( T \) is temperature in K, \( Z \) is the compressibility factor and

\[ M = 16.04 X_{\text{C}_1} + 30.07 X_{\text{C}_2} \]  \hspace{1cm} (6.2.17)

with \( X \) being the molar fractions of the components such that

\[ X_{\text{C}_1} + X_{\text{C}_2} = 1 \]  \hspace{1cm} (6.2.18)

**Oil density**

In-situ oil density is less easily calculated, due to the large variations in composition. If sufficient well data are available, oil properties can be calculated using reservoir engineering algorithms or from RFT data. Alternatively, for existing fields one can use PVT determinations from production reports.

For a first impression, oil density at surface and GOR can be converted to in-situ density using the relationships presented in Schowalter (1979).

\[ \rho_o = 1000((-0.00022 + 0.000002*(\rho_{\text{surface}}-10))/\text{(GOR-2000)+0.515}) \]  \hspace{1cm} (6.2.19)
Box 6.2.3. Leak dynamics

When the capillary displacement pressure of a water-wet top seal has been reached, it will leak hydrocarbons by flow through its interconnected porosity. The dynamics of two-phase flow through a water-wet capillary fault seal with a displacement pressure of 1.55 MPa were modeled using the one-dimensional reservoir simulator ONESIM (de Vries pers. comm. 2005). Rock properties are as shown, reservoir and seal have different permeabilities, porosities and capillary pressure curves, but have the same relative permeability and a common Leverett J curve. Figure 1 illustrates the model and some results.

Figure 1a is an illustration of sealing conditions at a fault where the imposed pressure difference is smaller than the capillary displacement pressure of the seal. At time = 0 water saturation equals 1 everywhere, and oil is injected from the left at constant pressure. The model is filled with oil up to the fault zone, and equilibrium is reached.

In figure 1b all parameters are the same as in the previous simulation, but the pressure difference across the fault is now higher than the displacement pressure. After the reservoir on the left is saturated with oil, an oil front moves through the seal. In steady state there is a relatively low oil saturation in the seal. The seal is leaking oil at a rate lower than expected from its bulk permeability only, owing to the low saturation.

Figure 1. ONESIM models of a fault which is a capillary seal at differential pressures below 1.55 MPa, but leaks at higher differential pressures. Pressure on the left is higher. At time = 0 the water saturation equals 1 everywhere, and oil is injected from the left. Both reservoir and seal are assumed to have the same relative permeability curves, and a common Leverett J curve. Oil and water viscosities are $10^{-3}$ Pa s. Simulations are run over 3000 years.
Figure 6.2.8. Typical mercury-air injection curves for sandstones and mudstones.

Figure 6.2.9.

(a) Illustration of the large-scale effects of capillary seal leakage during gas (green) and oil (red) migration, showing how in a carrier bed with variable capillary entry pressure different oil and gas accumulations evolve over time; after Schowalter (1981).

(b-e) Structural closure with a leakage path across a fault, showing how the interplay between rates of charge of oil and gas and the rates of leakage across the fault can produce widely different hydrocarbon accumulations. (green = gas, red = oil, yellow = water)
In this empirical formula, \( \rho_o \) is in-situ oil density (kg m\(^{-3}\)), \( \rho_{\text{surface}} \) is the density of the oil at surface (API) and GOR is the gas-oil ratio. The conversion from API units to kg m\(^{-3}\) is: \( \rho_o = \frac{141500}{(\text{API}+131.5)} \).

**Interfacial tension**

Accurate theoretical prediction of interphase tensions in hydrocarbon-water systems is a difficult and as yet unsolved problem. The two main parameters are density contrast and mutual solubility. Estimations are currently based on laboratory experiments but these are also problematic and sensitive to subtle changes in composition and interface structure. This has strong effect on the maximum columns calculated; interfacial tension is an important source of uncertainty in capillary seal calculations. Examples of interfacial tension data are presented in Schowalter (1979) and Watts (1987).

*Figure 6.2.10.*

Correlation between clay content, porosity and mercury-air Capillary entry pressure for a suite of Tertiary clays. After Hildenbrand et al. (2006)

*Figure 6.2.11.*

Envelope of mercury-air capillary curves from silty mudstones, muddy siltstones and mudstones of the Buntsandstein in the CEBS. Samples are from the depth range 800 – 4000 m.
Wetting properties

In parallel with the uncertainties in interfacial tensions, wetting is another poorly understood parameter in hydrocarbon-water-rock systems. Recent reviews are by Anderson (1986) and Swanson (1981). Note that wetting is a term related to contact angles in a solid-two fluid system, and does not directly relate to the fluid in contact with the rock at a given time.

An important point relevant to capillary seals is that clean silicate surfaces are water-wet. Silicate rocks which have been exposed to hydrocarbons, and other minerals (carbonates, anhydrite) may be mixed-wet or oil wet. In silicates this change is associated with the adsorption of a variety of compounds from crude oil (such as asphaltenes, see Anderson 1986).

6.2.3.4 (Hydro) fractured seal

By their capillary resistance, many seals can hold large hydrocarbon columns. From these, hydrocarbon leakage can occur by alternative mechanisms. Migration through fractures (tension fractures or shear fractures) is a likely candidate.

6.2.3.5 Brittle and Ductile Seals

Unfractured compacted mudrocks can hold very large hydrocarbon columns. In these rocks hydrocarbons can not move by Darcy flow through matrix porosity; flow through permeable fracture networks (besides diffusion through the pore fluid) is the most likely leak mechanism. Especially in the case of gas, a few thin fractures will allow depletion of a large field over geologic time. For this reason, ductility and its opposite, brittleness, are seen as key properties of effective seals.

In hydrocarbon seal studies, the terms brittle and ductile are used somewhat differently than in traditional rock mechanics. Because the key parameter here is the ability to transport fluids by one- or two phase flow, the following definitions are used (Ingram et al. 1997): a ductile shale is able to undergo plastic deformation without increasing its permeability (it may contain non-dilatant, impermeable fractures), whereas a brittle one will develop dilatant fractures and increase its permeability by many orders of magnitude when deformed (Fig. 6.2.14).

A brittle seal does not necessarily leak hydrocarbons: in the absence of tectonic deformation after the trap is filled, even very brittle lithologies can form excellent seals. Alternatively, initially dilatant faults may become re-sealed by processes such as the formation of a clay gouge (Holland et al. 2006) or cementation by hydrothermal fluids (Nollet et al. 2005). Therefore, an evaluation of the risk of seal leakage along fractures must always involve a structural and charge analysis.

Thus, a key to analysing shale top seals is to quantify the conditions where dilatant mechanical failure occurs.

This requires quantification of two competing parameters:

(i) mechanical properties of the formation under investigation, as given by its cohesion ($C$), friction angle ($\phi$)
and unconfined compressive strength (UCS), (a stronger rock is more brittle); (ii) the state of effective stress in the seal (high effective stress enhances ductility).

6.2.3.6 Fault Seals

Many hydrocarbon accumulations are affected in some way by the presence of faults in the reservoir and/or seal lithologies. In exploration and appraisal workflow, evaluation of the different leak pathways from a structure usually involves an assessment of the transport properties of faults. The two end-member cases here are flow along the fault (fault is completely open) and no flow in any direction (fault is completely sealing). The case in-between where there may be a seal along the fault but flow across the fault, or even more complicated systems of limited sealing over shorter distances in either direction, are the most complex. On the timescales of hydrocarbon production, the relative importance of sealing and non-sealing faults is quite different, and faults which have not played a role in the accumulation of oil or gas in a structure may form important flow barriers.

The first step in evaluating possible across-fault flow pathways consists of an analysis of possible juxtapositions across the fault. Although the principle can be illustrated in profile, the analysis is necessarily a 3D process, considering the topology of the fault system (Fig. 6.2.15) and the variation in offset along the fault. The final analysis is usually carried out using a fault plane diagram constructed in a plane parallel to the fault plane.

In the case of juxtaposition of permeable lithologies (Knipe et al. 1997) the transport properties of the fault gouge are the critical parameter controlling fluid flow. Prediction of the internal structure of fault gouges is difficult, due to the large amount of highly complex processes involved. The initially segmented geometry of fault zones and subsequent coalescence of segments, in combination with the contrast in mechanical and transport properties between the layers are the two most important factors in this evolution. The resulting ranges of structures are illustrated in figure 6.2.16.

It is clear from the above that fault zones can have vastly higher or lower permeabilities than their country rocks, and thus form either barriers or conduits for fluid flow. The amount and type of clay in a fault is one of the main parameters controlling the mechanical strength and fluid transport properties of the fault gouge (Logan and Rauenzahn 1987; Lupini et al. 1981; Heynekamp et al. 1999; Rawling et al. 2001). Much of the current re-

Figure 6.2.13. Mercury–air capillary curves from Tertiary claystones in the CEBS

![Figure 6.2.14. Illustration of the top seal above an oil accumulation, which becomes deformed after charge. In the first case the seal is ductile and can deform without becoming permeable: the accumulation is not affected. In the second case the deforming seal becomes permeable, and the accumulation moves to a higher structure](image)
search on fault seal is therefore focused on the process of clay being incorporated into the fault gouge, often called “clay smear”.

Clay smear (Fig. 6.2.17) is a loosely defined term born in hydrocarbon geology (van der Zee 2002; Aydin and Eyal 2002); its usage differs between publications and the definition of processes operating is often unclear. In the most general meaning, the term includes all processes which somehow transform clay in the wall rock into clay which is part of the fault zone. Processes included are clay abrasion (Lindsay et al. 1993), shear in releasing fault links (Bashir Koledoye et al. 2000), preferred smear and lateral clay injection (Lehner and Pilaar 1997; van der Zee et al. 2003).

For applied studies of clay smear evaluation a number of semi-empirical tools are available (Bouvier et al. 1989; Fulljames et al. 1997; Lindsay et al. 1993; Fristad et al. 1997). An overview and comparison of these methods is given by Yielding et al. (1997). Most of these methods are based on the assumption that the fault gouge is a re-worked equivalent of the wall rock without addition or removal of material, and therefore has, on average, the same clay fraction. Unfortunately this assumption is statistical in nature, and it does not specify which of the different structures shown in figure 6.2.16 is present. Therefore, predictions based on this method have a high degree of uncertainty because rock properties are not included in the calculation.

A simplified model in Mohr space

To illustrate the main mechanical aspects of clay injection, we present a simple model using Mohr diagrams (Fig. 6.2.18).

The structural evolution is at a stage where the fault has just formed across the whole area considered, but the fault throw is still very small and the two sides of the
fault in the releasing bend have been partially unloaded but have not yet separated. We now consider a point (A) in the releasing bend. The maximum principal effective stress ($\sigma'_1$) approximately equals the overburden load. The minimum principle effective stress ($\sigma'_3$) in this point is approximately horizontal, and is progressively reduced by the fault movement. In Mohr space this results in the Mohr circle with its right leg fixed and its left leg moving to the left. At this point there are two possibilities: (i) The yield envelope is reached before $\sigma'_3 = 0$, and plastic flow of the clay towards the pull apart structure is initiated. (ii) In a more cohesive clay, the yield envelope is not reached before $\sigma'_3 = 0$. The pull-apart structure will open because the clay layer can support the vertical load without becoming plastic. Therefore, there will be no injection, but the formation of a dilatant void instead. In this location other processes may take place such as vein growth (Hilgers et al. 2004).

For the case of a linear Mohr-Coulomb plasticity criterion the condition for the onset of lateral injection can be written by:

$$ C = \sigma'_1 \left(1 - \sin \phi \right) / \left(2 \cos \phi \right) $$

(6.2.26)

Here $C$ is cohesion (MPa), $\sigma'_1$ is the maximum principle effective stress (MPa) and $\phi$ is internal friction angle (degrees). Equation 6.2.26 is a first-order description of the conditions required for injection of clay. It contains a description of the interdependence of in-situ stress and rock strength. The usefulness of this equation comes from the fact that all the parameters involved can be in principle obtained from wire line log and cuttings data. Therefore, this relationship can be used to predict the onset of clay injection in the subsurface.

6.2.3.7 Numerical modelling of petroleum flow

The flow of petroleum in sedimentary systems is dominantly governed by the interplay of forces which on the one hand drive and on the other resist movement. Petroleum generated in mature source rocks migrates within these usually fine-grained rocks until it reaches more
porous and permeable carrier rocks. These two closely linked processes are commonly referred to as “primary migration” and “expulsion” (Welte et al. 2000). Petroleum flow within the permeable carrier rocks until a trap is reached is referred to as “secondary migration”. Long- and short-term leakage processes out of reservoir structures through the seal (capillary leakage), along fault and fracture systems or past “spill points” are referred to as “tertiary migration” (Sylta 2004).

The main driving force for migration is the buoyancy of the petroleum phase, as a result of its lower specific density compared to surrounding pore water under subsurface conditions. Overpressure gradients in the continuous pore water phase can add to this flow drive (Hubbert 1953). Viscous and capillary forces resist this driving force. The magnitude of the viscous forces is controlled by the dynamic viscosity of the petroleum phase and the relative permeability of the rock matrix with respect to this phase. The capillary forces manifest themselves in terms of the capillary entry pressure which depends on the wettability properties of the pore system and interfacial tension between the fluid phases. Fluid flow will occur whenever the driving force exceeds the resisting force.

Modern basin simulation software packages use one or more of the four following concepts to model fluid flow under subsurface conditions on the geological time scale:

- full Darcy flow modelling
- Ortho-contouring
- Ray-tracing
- Invasion percolation

The multiphase Darcy flow law is the most commonly accepted concept to achieve full physical modelling of fluid flow. The Darcy equation describes the relationship between fluid flow, permeability and pressure gradients (potential gradients) in porous media (e.g., Bear 1972). The Darcy flow concept assumes that flow velocities are controlled by permeability and fluid viscosity functions and it is therefore considered the best way to describe the rates of fluid flow in a physically correct deterministic fashion. The multi-component formulation of these concepts can take into account the dissolution–exsolution behaviour of individual chemical components/compound classes into and out of the phases. This is an important prerequisite because pressure and temperature conditions

Figure 6.2.16. An illustration of the main structural elements found in fault zones in sediments (after van der Zee et al. 2003). (a) heterogeneous simple shear deformation, without strong localisation in a discrete fault plane; (b) heterogeneous simple shear, with lateral transition into a sharply localised zone of deformation; (c) lens structure between two branches of a fault plane, usually showing a higher degree of deformation than the surrounding rocks; (d) fault gouge consisting of parallel strands, with telescoping of the individual layers; (e) fault gouge with lateral injection of a weak layer (this paper); (f) fault gouge formed by the process of preferred smear of the soft layer; (g) fault gouge formed by disruption of the stronger layer followed by mixing of the fragments; (h) the brittle end member fault, with open fractures, developing preferentially in releasing sections.
usually vary during secondary migration. This method is probably the best when it comes to dealing with the dynamics of petroleum migration, including, in particular, the time required for migration. However, it is also the calculation method which is computationally most demanding. Therefore, several approaches have been investigated to simplify simulation algorithms.

Ortho-contouring

The simplest method of hydrocarbon flow modelling is ortho-contouring, which focuses on assessing the principal flow directions rather than explicitly modelling the fluid flow process. Contours are constructed at angles of 90 degrees to the strike of a carrier top map. The method is attractive to geologists because it can be performed both manually and by computerised methods. It neglects, however, effects of hydrodynamics (overpressure variations) and carrier entry pressure variations (e.g., carrier facies changes). This method does not provide any information on fluid flow volumes or masses and should therefore not be used for detailed flow analysis.

Ray-tracing

The ray-tracing method constructs migration flowpaths based on the top depth map of a carrier bed (Sylta 1993). The underlying assumption is that hydrocarbon flow rates and therefore hydrocarbon saturations change very slowly as compared to the duration of simulation time-steps and that migration can therefore be modelled as a steady state process. The thickness of the migration pathways is calculated from the flow rates, carrier bed dips and permeability. Ray-tracing methods are typically applied to migration within highly permeable carrier beds (Skjervøy and Sylta 1993). The method operates on high-resolution grids and requires only short computation times.

Invasion percolation

In the invasion percolation method, flow is modelled based on a pore “entry pressure” (threshold pressure) and petroleum saturation is often assumed to be constant. Migration between adjacent cells is computed by searching for the neighbouring cell with the lowest resistance to flow. All flow is then modelled to occur into this neighbouring cell. Repeated application of the algorithm results in the identification of a “backbone” along which essentially all migration of the non-wetting fluid (petroleum) occurs. This method is extremely efficient due to the concentra-

Figure 6.2.17. Normal fault in a deltaic sand-mudstone sequence, airport road outcrop, Sarawak, Malaysia. Note the continuous seam of clay in the fault gouge, making this section of the fault a good lateral seal. However, the processes which led to the formation of this clay smear are not clear.

Figure 6.2.18. Schematic sketch of the stress evolution, in Mohr space, of a point (A) in the clay layer in a releasing bend of a normal fault.
tion of computing resources on the calculation of the hydrocarbon migration backbone. The main advantage of the method is the computational speed which makes it possible to create transport simulations on basin-scale models having seismic property resolutions.

All four approaches outlined above are currently in use in commercial basin modelling packages. Applications have, however, shown that in some cases one of these methods tends to provide better results (in terms of reproducing observations) than the others. Combined (“hybrid”) methods, e.g., with Darcy flow assumed in low permeability rocks and ray-tracing in carrier systems, are also available.

Expulsion of petroleum from source rocks is a slow process. Expulsion rates vary between a few grams to a few kilograms of hydrocarbons per m$^3$ source rock per million years. If the rate of build-up of the driving force (i.e., an overpressure gradient towards the resisting pores) is very low, it can be shown that viscous forces are negligible. Then, the fluid flow is dominated by capillary forces and the result is an extremely focused flow with very narrow pathways in sandstones (<1 m). These arguments have led England et al. (1987) to describe the fluid flow pattern as “migration ganglia” or “rivers of oil (and gas)”, describing the very focused and discrete distribution of hydrocarbons in sedimentary basins.

6.2.4 Geological aspects of fluid transport

6.2.4.1 Fluid flow regimes

Different flow regimes occur in sedimentary basins; meteoric, compactional and thermobaric (Fig. 6.2.19). In addition, thermal or thermohaline convection and seismic pumping may act as driving mechanisms for fluid flow in sedimentary basins (Bjørlykke 1993; Garven 1995; Harrison and Tempel 1993; Sharp et al. 2001; Sibson et al. 2003; Wood and Hewett 1984; Gaupp et al. this volume).

**Meteoric water flow** is a consequence of the infiltration of rainfall into the subsurface. Its direction and intensity depends on the topography (hydraulic head) and the permeability, lateral extension and connectivity of the conducting rock layers (aquifers). Therefore, meteoric fluid flow is most important at shallow depths or in uplifted parts of a basin. At topographic highs along marine basins, however, meteoric waters can displace connate fluids down to depths of 3 km (Harrison and Temple 1993). For example, freshwater springs discharge at depths of several hundred meters beneath sea level in many coastal regions.

Some meteoric water systems extend over large distances and transport considerable amounts of water over geological time. Under varying pressure and temperature conditions the solubility of ionic and gaseous species in the water changes. This may result in a redistribution of these species within the hydrological system. Cramer et al. (1999) investigated the potential role of the aquifer transport of dissolved methane in the formation of the giant Urengoy gas field in Western Siberia. Assuming that methane saturation of the water in the Cretaceous aquifer of the West Siberian Basin occurs some 500 km south of the Urengoy field, they concluded that long-range migration in combination with a Neogene uplift and gas exsolution can account for the volumes of isotopically light methane in the giant gas field (Fig. 6.2.20).

If overburden pressure increases faster than compaction, overpressure can form in under-compacted zones. **Compactional flow** occurs when pore fluid is expelled
from the over-pressured zone. As pore fluids move from compaction of fine-grained sediments into more permeable coarser-grained sediments, overpressures will dissipate within relatively short geological times and the hydrostatic pressure regime is approached. Consequently, compaction-driven flow is generally considered only as a localized process and is of lesser importance in regional fluid movement. The compactional flow regime reaches from the surface down to a depth of approximately 4 km (Harrison and Temple 1993).

Thermal gradients increase at greater burial depth and can induce thermal convection through permeable strata. In particular in areas with dominant salt structures, which are excellent heat conductors, the temperature field becomes disturbed. In addition, salinity gradients can occur in the vicinity of salt domes triggering thermohaline convection as described by Sharp et al. (2001) or Magri et al. (this volume), which can drive fluid flow through large portions of the sedimentary section. Fluid convection not only transports heat and energy but also dissolves species which drive diagenetic processes. Repetitive cycling of brine within an essentially closed system may result in considerable redistribution of matter by dissolution and precipitation/exsolution processes. Because such convective flow can develop on the scale of entire basins, it is an important process in diagenesis and emplacement of petroleum.

In the deepest portion of sedimentary basins, thermobaric flow can occur. Fluids move in response to pressure changes generated by the release of mineral-bound water, the generation of hydrocarbons, increased heat flow and/or continuous lithostatic loading (Galloway 1984). The dehydration of clay minerals like the illitisation of smectites can release significant amounts of water to the pore fluid. Increased heat flow may cause volume expansion of pore water also raising static pressure. However, low volumes and permeabilities at great depth restrict water movement over geological time spans (Galloway 1984) and small fluid volume movements rapidly dissipate the overpressure generated in the thermobaric regime.

Sibson (2003) defined the term seismic pumping as fluid flow driven by repeated episodes of dilation and contraction near and within seismically active fault zones. Dilation drives fluids towards newly opened cracks and dilated pore space. Because of the low permeability of basement rocks, groundwater migrates slowly to faults, resulting in continued fluid flow for weeks following an earthquake.

Within sedimentary basins different pressure compartments separated by impermeable barriers preventing pressure re-equilibration can exist. Big lateral pressure differences occurring across faults (Mann and Mackenzie 1990) indicate that faults can act as a lateral seal for the pressure compartments. Over- and underlying sealing layers act as vertical boundaries for the pressure compartments (Borge 2002; Wangen 2001a,b). Distinct pressure-depth gradients in the pre-Cretaceous sandstones in the Viking and the Central Graben suggest pressure compartmentalisation of the sandstones (Darby et al. 1996; Moss et al. 2003; di Primio et al. this volume). Osborne and Swarbrick (1997) evaluated the different mechanisms for generating overpressure in sedimentary basins and concluded that disequilibrium compaction is the most feasible overpressuring mechanism in thick and rapidly deposited sediments.
mudstone sequences. This is supported by the observation that overpressured fluids are often found in young sedimentary sequences that are characterised by rapid sedimentation of thick shale-sand successions (Harrison and Summa 1991).

No consensus exists on concepts of static and dynamic conditions in overpressured systems (Moss et al. 2003). The static school considers overpressure as a static system of isolated fluid compartments being permanently sealed off from pressure communication with neighbouring rocks (Bradley 1975; Powley 1990; Hunt 1995; Bradley and Powley 1994). Alternatively, the dynamic school (Bethke 1986; Bredehoeft et al. 1994; Neuzil 1994) considers overpressure formation a transient phenomenon which results from fluid flow in a dynamic, hydraulically continuous system (Moss et al. 2003). Therefore, fluid flow is never completely sealed off and overpressure bleeds off on geological time scales of million of years (Deming 1994; Neuzil 1994).

Fluid flow and diagenesis

Aqueous phase transport of dissolved chemical species through the pore space of sedimentary rocks may cause cement precipitation or leaching of mineral grains. Diffusive transport is usually relevant only over short distances and/or over long time spans. With respect to their chemical composition, different fluid environments and sources can be distinguished: meteoric water, formation water and fluids released from mineral bound water.

Meteoric water is in general an acidic fluid with low salinity which infiltrates the rocks from the surface and reacts with the minerals. It replaces or mixes with existing connate water and thus changes the chemical composition of the fluid and initiates early diagenetic alteration of the rocks. The solute load of the fluid increases with geological time. In the North Sea region, meteoric fluid flow is commonly associated with shallow depth and early diagenetic processes (Haszeldine et al. 2000). Dissolution of K-feldspar and the formation of kaolinite is frequently interpreted as an indicator of inflow of meteoric water into sandstones (e.g., Bjørklykke 1994; Brosse et al. 2003). The kaolinitisation of K-feldspar is not an isochemical reaction because potassium and silica form as reaction products that have to be removed by pore fluid flow if this process is to proceed.

At greater depth or when the temperature field is disturbed by geological structures such as diapirs, thermal convection can transport fluids through the rock, moving them through changing temperature fields or bringing them into contact with rocks of different mineral composition. The new conditions change the existing chemistry of the fluid rock system which will cause mineral reaction thus driving the diagenetic evolution of the rock.

The capacity for solute transport by compaction driven flow is generally very low and the resulting effect on diagenetic processes is limited. If, however, compactional flow is focused by faults this can allow an efficient transfer of overpressured fluids of distinctively different composition into regions of lower temperature and normal pressure which may lead to significant mineral precipitation near the fracture zone.

A study of the diagenetic evolution of Permian Sandstones in North Germany (Gaupp et al. 1993; Platt 1993) indicates that K-feldspar dissolution and illite formation are closely linked to fault zones indicating that faults acted as conduits for fluid flow which triggered subsequent mineral reactions. A study of the diagenetic evolution of Gulf Coast mudrocks (Land and Milliken 2000) shows a loss of silica and carbonate and an import of potassium. No clear answer is given with respect to the relevant transport mechanism but Land and Milliken (2000) consider diffusive transport unlikely because chemical gradients in the large-scale sand-mud system are insufficient to drive diffusion. Alternatively, thermal convection or transport along fractures are considered possible but unproven transport processes.

Although there is some controversy as to the exact extent and efficiency of the different driving mechanisms for fluid flow in sedimentary basins (e.g., Haszeldine et al. 2000; Bjørklykke 1993; Land and Milliken 2000), fluid flow is an efficient process for transporting reactants through the pore space of rocks. It is, therefore, an absolutely essential process which drives diagenetic processes in sedimentary basins. Just which transport process, fluid flow or diffusion, controls diagenesis depends on the geological conditions of the system and it may change over time.