8 Hydraulic Properties of Rocks

8.1 Basic Concepts and Terminology

Hydraulic properties of water bearing formations are important as they govern their groundwater storage and transmitting characteristics. These are described below.

Porosity (η) Porosity, η is a measure of the interstices or voids present in a rock formation. It is defined as the ratio of volume of voids, V_v to the total volume, V of the rock mass (Eq. 8.1)

$$\eta = \frac{V_V}{V} \tag{8.1}$$

Porosity, (η) is given as a percentage or as decimal fraction. Porosity is of two types, primary and secondary. Primary porosity is the inherent character of a rock which is developed during its formation. Secondary porosity is developed subsequently due to various geological processes, viz. fracturing, weathering and solution activity. In unconsolidated rocks, primary porosity is of importance but in hard rocks secondary porosity is of greater significance.

Porosity is controlled by: (a) shape and arrangement of constituent grains, (b) degree of sorting, (c) compaction and cementation, (d) fracturing, and (e) solution. The geometrical arrangement of constituent grains (packing) and sorting have important influence on porosity. Well sorted clastic material has high porosity irrespective of grain size. In poorly sorted material, porosity is less as small-size grains occupy pore spaces between bigger grains. Compaction and cementation reduces porosity. In unconsolidated formations, porosity at deeper levels will be less due to compaction, e.g. shales have lower porosity than clays.

The fractured rock formations are made up of two porosity systems, (a) the intergranular porosity or matrix porosity (η_m), formed by intergranular void spaces, and (b) the secondary porosity developed due to fractures and solution cavities, termed as fracture porosity, (η_f). Therefore, in fractured rocks, the total porosity is the sum of matrix and fracture porosities, i.e.

$$\eta = \eta_m + \eta_f \tag{8.2}$$

The two porosities can be expressed in the conventional manner as

$$\eta_m = \frac{Matrix\ void\ volume}{Total\ bulk\ volume}$$
$$\eta_f = \frac{Fracture\ void\ volume}{Total\ bulk\ volume}$$

In laboratory, porosity of rock samples can be estimated by immersion in liquids, density test and by gas-porosity meters (UNESCO 1984b). In field, geophysical logging methods, viz. resistivity, neutron and gamma methods can be used for determining porosity.

In the field, fracture porosity can be estimated from scan line method by the relation η_f=Fa, where F is the number of joints per unit distance intersecting a straight scan line across the outcrop and a is the mean aperture of fractures. The porosity of natural materials may range from almost zero in hard massive rocks to as much as 60% in clays (Table 8.1).

Laboratory measurements on a variety of fractured rocks show that fracture porosity, η_f is considerably less than the matrix porosity. η_m is reported to generally vary from 0.1% to 8% and n_f from 0.001% to 0.01% (Lee and Farmer 1993).

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Table 8.1 Representative values of porosity (η), specific yield (S_y) and specific retention (S_r) of geological materials. (After Morris and Johnson 1967; Hamill and Bell 1986)

Geological formation	η(%)	S_y(%)	S_r(%)
Unconsolidated deposits			
Gravel	28–34	15–30	3–12
Sand	35–50	10–30	5–15
Silt	40–50	5–20	15–40
Clay	40–60	1–5	25–45
Dune Sand	40–45	25–35	1–5
Loess	45–50	15–20	20–30
Rocks			
Sandstone	15–30	5–25	5–20
Limestone, dolomite	10–25	0.5–10	5–25
Shale	0–10	0.5–5	0–5
Siltstone	5–20	1–8	5–45
Till	30–35	4–18	15–30
Dense crystalline rock	0–5	0–3	–
Fractured crystalline rock	5–10	2–5	–
Weathered crystalline rock	20–40	10–20	–
Basalt	5–30	2–10	–

Void Ratio (e) This is generally used in soil mechanics and is expressed as

$$e = \frac{V_v}{V_s} \tag{8.3}$$

where V_S is volume of mineral grains and V_v is as defined earlier. Void ratio has large numerical variation. In natural soils, where total porosity ranges from 0.3 to 0.6, the corresponding void ratio range is 0.45–1.5. The relation between total porosity, η and void ratio e, can be expressed as

$$\eta = \frac{e}{1+e}, \quad \text{or} \quad e = \frac{\eta}{1-\eta} \tag{8.4}$$

Specific Retention (S_r) This is a measure of the volume of water retained by the rock material against gravity on account of cohesive and intergranular forces. It can be expressed as

$$S_r = \frac{V_r}{V} \tag{8.5}$$

where V_r is the volume of water retained. Specific retention depends on the *specific surface* of constituent mineral grains which in turn is influenced by the grain size, shape and type of clay minerals present. Specific surface is defined as the area per unit weight of the material and is expressed in $m^2 g^{-1}$. The specific surface values of coarse grained material, such as gravel and sand, is small compared with silt and clay size fractions. Among clay minerals, non-swelling clays have specific surfaces in the range of 10–30 $m^2 g^{-1}$, but swelling clays such as montmorillonite have large values of about 800 $m^2 g^{-1}$. A similar property is the specific surface area, (S_{Sp}), defined by

$$S_{SP} = \frac{\text{Total surface area of the interstitial voids}}{\text{Total volume of the medium}}$$

S_{sp} has the dimensions of L^{-1}. In fine-grained material, S_{sp} will be more, viz. in sands, S_{sp} will be of the order of $1.5\times10^4\,m^{-1}$ but in montmorillonite, it is about $1.5\times10^9\,m^{-1}$ (de Marsily 1986). These properties are of importance in the adsorption of water molecules and ions on the surfaces of mineral grains, especially on clay minerals.

Specific Yield (S_y) This is defined as the ratio of the volume of water that an unconfined aquifer will release from storage by gravity, to the total volume of fully saturated aquifer material. It is expressed either as a decimal fraction or as a percentage. Specific yield depends on the duration of drainage, temperature, mineral composition of water, grain size and other textural characteristics of aquifer material. Values of specific yield (S_y) of some common rock materials are given in Table 8.1.

Effective Porosity (η_e) Effective porosity or *kinematic porosity* is the same as specific yield (de Marsily 1986). The concept of effective porosity indicates that all the pores do not participate in the flow of water. Fine grained and poorly sorted materials have low effective porosity as compared with coarse grained and well sorted material, due to the greater retention of water on account of intergranular forces. Instead of total porosity, the effective porosity (η_e) is more important for estimating the average velocity of groundwater and transport of contaminants as discussed in Sects. 7.1.2 and 7.4.1.

In crystalline and other hard rocks, the size and interconnection of fractures are mainly responsible for imparting effective porosity to the rock mass. In such rocks, although total porosity may be high but due to unconnected fractures, the effective or kinematic porosity will be less, viz. in granites and other crystalline rocks, although total porosity may be 1–10% but the effective porosity is very small ($5 \times 10^{-5} - 1 \times 10^{-2}$). Similarly, in dolomites, which are formed as a result of diagenesis, although the rock may acquire high secondary porosity due to reduction in volume of mineral grains, but effective porosity will be less.

Hydraulic Conductivity (K) This is a measure of the ability of a formation to transmit water. It has the dimensions of LT^{-1} and is usually expressed in $m\,s^{-1}$.

In terms of Darcy's law (Eq. 7.12), K can be expressed as

$$K = \frac{V}{dh/dl} \tag{8.6}$$

where K is hydraulic conductivity, V is groundwater velocity, and dh/dl is hydraulic gradient.

In the USA, K was earlier expressed in two forms– as the Meinzer's (Laboratory) coefficient of permeability, K_m and also Field coefficient of permeability (K_f), both using units of gal $d^{-1}\,ft^{-2}$. The main difference between the two being that K_m is expressed at a constant temperature of 60°F (15.6°C) while K_f is measured at the actual field temperature. As field conditions do not influence the groundwater temperatures to any significant extent, the distinction between K_m and K_f has now been discarded.

The hydraulic conductivity depends both on the properties of the medium (rock material) as well as of the fluid. In sedimentary formations, grain size characteristics are most important as coarse grained and well sorted material will have high hydraulic conductivity as compared with fine grained sediments like silt and clay. Increase in degree of compaction and cementation reduces hydraulic conductivity. In hard (fractured) rocks, K depends on density, size and inter-connection of fractures (Sect. 8.2).

Permeability (k) It is a more rational concept than hydraulic conductivity (K) as it is independent of fluid properties and depends only on the properties of the medium. Fluid properties which influence hydraulic conductivity are viscosity (μ), expressing the shear resistance, and specific weight, (γ), expressing the driving force of the fluid.

The relation between hydraulic conductivity, (K) and properties of the medium and the fluid can be expressed as

$$K = \frac{cd_e^2\,\gamma}{\mu} = \frac{k\gamma}{\mu} \tag{8.7}$$

where c is a dimensionless constant also known as shape factor and d_e is effective grain size; c depends on porosity and packing etc. Equation 8.7 when substituted in Darcy's equation gives

$$k = \frac{\mu\, Q/A}{\gamma\, \left(dh/dl\right)} \tag{8.8}$$

The value of k can be given in darcy units which in terms of Eq. 8.8 is expressed as

$$1\ darcy = \frac{\left[1\ \left(cm^3/sec\right)/cm^2\right]\ 1\ cP}{1\ atm/cm} \tag{8.9}$$

Thus, a porous saturated medium will have a permeability of one darcy if a fluid of 1 centipoise (1cP) viscosity will flow through it at a rate of $1\,cm^3\,s^{-1}$ per cm^2 cross-sectional area under a pressure or equivalent hydraulic gradient of $1\,atm\,cm^{-1}$.

According to Eq. 8.9, k has the units of area. As the value of k is very small, it is also expressed in square micrometres $(\mu m)^2$.

$$1(\mu\mathrm{m})^2 = 10^{-12}\mathrm{m}^2$$

By substitution of appropriate units in Eq. 8.9, it can be shown that,

$$\begin{aligned} 1\mathrm{darcy} &= 0.987(\mu\mathrm{m})^2 = 10^{-8}\mathrm{cm}^2 \\ &= 10^{-5}\mathrm{ms}^{-1}(\mathrm{approx.}) \end{aligned}$$

The range of values of hydraulic conductivity, K and permeability, k are given in Table 8.2. Conversion factors for the various common units of K and k are given in Appendix. It could be noted from Table 8.2 that the permeability of natural materials has wide variation. As the permeability of crystalline rocks and other tight formations is usually very small, k in such cases is usually expressed in millidarcy (md) which is approximately equal to $10^{-8}\,m\,s^{-1}$. The permeability of dense unfractured rocks is usually very low being up to $1\,md^{-1}$, and usually below $0.001–0.5\,md^{-1}$. Fractures increase the permeability by several orders of magnitude above the

Table 8.2 Range of values of hydraulic conductivity and permeability for various types of geological materials

Hydraulic conductivity, K (ms^{-1})	1	10^{-1}	10^{-2}	10^{-3}	10^{-4}	10^{-5}	10^{-6}	10^{-7}	10^{-8}	10^{-9}	10^{-10}	10^{-11}	10^{-12}	10^{-13}
Permeability, k (darcy)	10^5	10^4	10^3	10^2	10	1	10^{-1}	10^{-2}	10^{-3}	10^{-4}	10^{-5}	10^{-6}	10^{-7}	10^{-8}
Relative values		Very high		High		Moderate				Low			Very low	
Representative materials														
Unconsolidated deposits														
Gravel	←	—	—	→										
Clean sand			←	—	—	→								
Silty sand					←	—	—	→						
Clay till (often fractured)							←	—	—	—	→			
Rocks														
Shale & siltstone (unfractured)											←	—	→	
Shale & siltstone (fractured)								←	—	—	→			
Sandstone							←	—	—	→				
Sandstone (fractured)					←	—	—	—	→					
Limestone & dolomite								←	—	—	→			
Karst limestone & dolomite					←	—	—	→						
Massive basalt								←	—	—	→			
Vesicular & fractured basalt					←	—	—	—	→					
Fractured & weathered crystalline rock						←	—	—	—	→				
Massive crystalline rock										←	—	—	—	→

solid rock mass. The fracture permeability can be 100 and even 1000 md^{-1}.

Transmissivity (T) This is defined as the rate of flow of water at the prevailing field temperature under a unit hydraulic gradient through a vertical strip of aquifer of unit width and extending through the entire saturated thickness of the aquifer (Fig. 8.1). Transmissivity (T) has dimensions of L^2T^{-1} and is expressed in $m^2 d^{-1}$ or $m^2 s^{-1}$. Darcy's law, in terms of T, can be written as

$$Q = TIL \tag{8.10}$$

where Q = rate of flow, I = hydraulic gradient, L = width of the flow section, measured at right angles to the direction of flow.

In confined aquifer, T = Kb, where b is the saturated thickness of the aquifer. In unconfined aquifer, the saturated thickness will be less than the true thickness. Here it is assumed that K is isotropic and constant across the thickness of the aquifer which may be horizontal or dipping. Transmissivities greater than 1000 $m^2 d^{-1}$ represent good aquifers for groundwater exploitation. In geothermal reservoir, transmissivity is usually expressed in terms of permeability-thickness (kb) in units of d-m ($1\,m^3 = 10^2$ d-m). We return to this subject with respect to geothermal reservoirs in Chap. 18.

Storativity (S) Storativity of an aquifer is defined as the volume of water which a vertical column of the

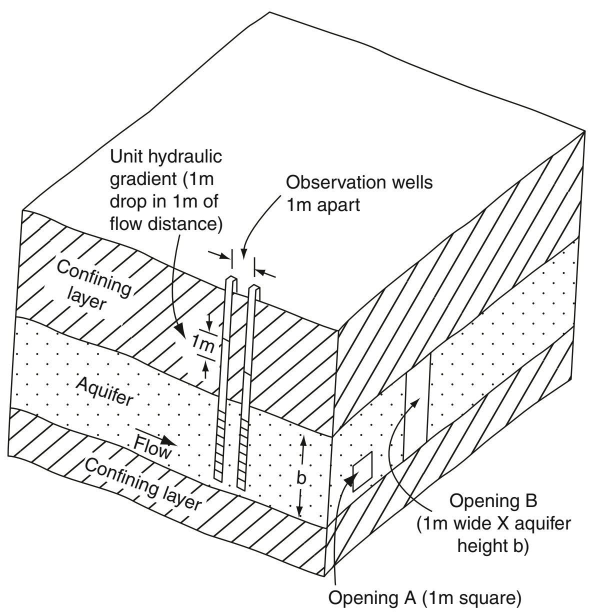

Fig. 8.1 Diagram illustrating coefficients of hydraulic conductivity (K) and transmissivity (T). Flow of water through opening A will be equal to K and that through opening B equal to T

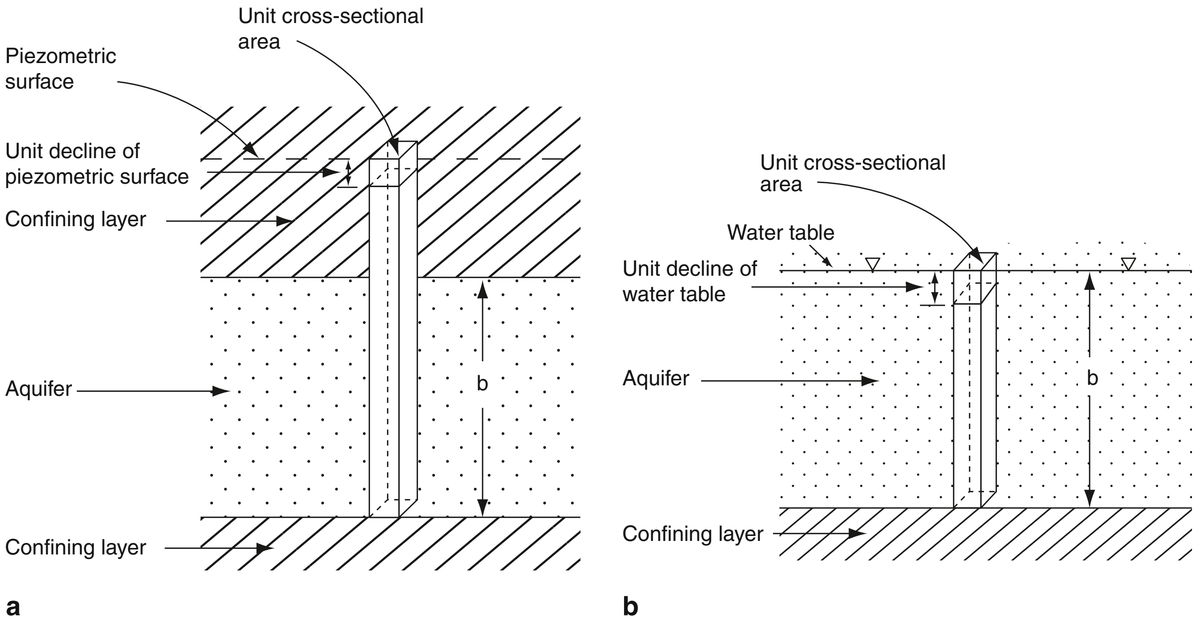

Fig. 8.2 Diagrams illustrating coefficient of storage (S) for: **a** confined aquifer, and **b** unconfined aquifer. (After Ferris et al. 1962)

aquifer of unit cross sectional area releases from storage as the average head within this column declines by a unit distance (Fig. 8.2). It is therefore dimensionless. In a confined aquifer, where water released from or taken into storage is entirely due to compressibility of aquifer and water, the storage coefficient is given by $S = bS_s$, where S_s is the specific storage, defined later. Value of storativity in confined aquifer is of the order of 10^{-3}–10^{-6}. In an unconfined aquifer, storativity S, is given by $S = S_y + bS_s$. Usually $S_y >> bS_s$, thus storativity of unconfined aquifer for all practical purposes is regarded equal to its specific yield or effective porosity, (η_e) (Hantush 1964). Storativity in unconfined aquifers ranges from 0.05 to 0.30.

The relation between storativity and the compressibility of the aquifer material and of water can be expressed as

$$S = \gamma \eta b \left(\beta + \frac{\alpha}{\eta} \right) \tag{8.11}$$

where, β is the compressibility of water (4.7×10^{-10} Pa^{-1}), α is the compressibility of solid skeleton of the aquifer and S, η and b are defined earlier.

Specific Storage (S_S) This is the volume of water which a unit volume of the confined aquifer releases from storage because of expansion of water and compression of the aquifer under a unit decline in the average hydraulic head. It has the dimension of L^{-1}. S_s is used exclusively in confined aquifer analysis, as in unconfined aquifer the water released from storage is mainly due to gravity drainage and not due to the compressibility of aquifer material or of water. S_s is a more fundamental parameter as compared to S as the latter depends upon both the specific storage and the aquifer geometry. In terms of Eq. 8.11

$$S_s = \gamma \ (\alpha + \eta\beta) \tag{8.12}$$

Therefore, S_S depends both on coefficient of compressibility (α) and porosity (η) of the rock. As hard dense rocks have low porosity and low coefficient of compressibility, S_S is also less as compared with sands and clay formations. Some representative values of α and S_S, are given in Table 8.3.

Table 8.3 Range in values of the coefficient of compressibility of solid material (α) and specific storage (S_s). (After Domenico 1972; Streltsova 1977)

Rock type	α (Pa^{-1})	S_s (m^{-1})
Dense rock	10^{-12}–10^{-10}	10^{-7}–10^{-5}
Fissured and jointed rock	10^{-10}–10^{-9}	10^{-5}–10^{-4}
Sand	10^{-9}–10^{-8}	10^{-4}–10^{-3}
Clay	10^{-8}–10^{-7}	10^{-4}–10^{-2}

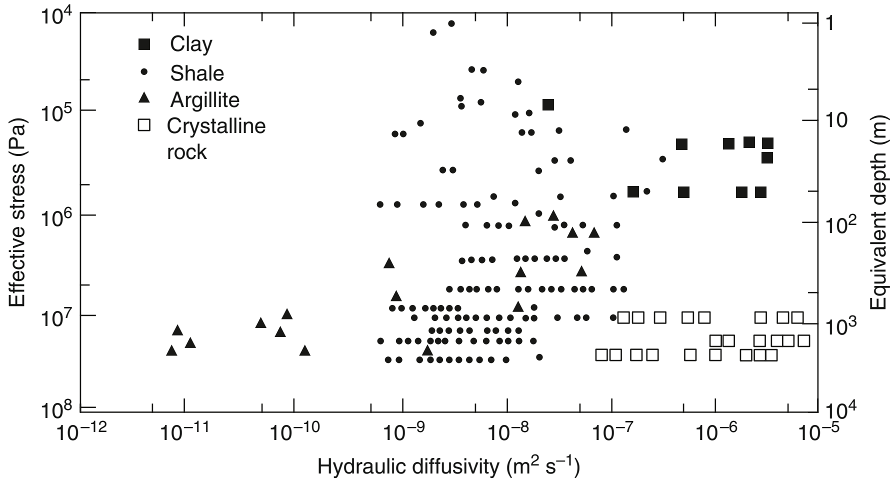

Fig. 8.3 Measured values of hydraulic diffusivity (κ) for a variety of low permeability rocks plotted against effective stress and the equivalent depth. (After Neuzil 1986)

Hydraulic Diffusivity (κ) This is a single formation characteristic that couples the transmission properties, K and storage property, S_s or alternately T and S.

$$\kappa = T/S = K/S_s \tag{8.13}$$

k has dimensions of L^2T^{-1}. It is a significant property of the medium for transient flow and has a major influence on the drawdown response around a pumped well. A comparison of measured κ in rocks of low permeability under varying stress is illustrated in Fig. 8.3. The equivalent depth is also given. It could be noted that in argillaceous materials, the hydraulic diffusivity, κ is generally less (10^{-9}–$10^{-7}\,m^2\,s^{-1}$) than in crystalline rocks, viz. gabbros and granites (10^{-7}–$10^{-5}\,m^2\,s^{-1}$), which appears to be due to the low S_s of crystalline rocks.

Leakage Coefficient or Leakance (Dimensions T^{-1}) This is the property of the semi-confining (aquitard) layer. It is equal to K^1/b^1, where K^1 and b^1 are the vertical hydraulic conductivity and thickness of the aquitard respectively.

Hydraulic Resistance (C) It is the reciprocal of leakage coefficient. It indicates resistance against vertical flow in an aquitard. It is equal to b^1/K^1 and has the dimensions of time. If hydraulic resistance, C = ∞, the aquifer is confined.

Leakage Factor (B) This determines the distribution of leakage through an aquitard into a leaky aquifer. It is defined as

$$B = \sqrt{TC} \tag{8.14}$$

Leakage factor has dimensions of length and is expressed in metres. High values of B indicate greater resistance of the semi-previous strata to leakage.

Boulton's Delay Index (1/α) This is a measure of the delayed drainage of an unconfined aquifer. It has the dimensions of time. The value of 1/α may vary from about 50 min in coarse sand to 4000 min in silt and clay.

Drainage Factor (D) Drainage factor ($D = \sqrt{T/\alpha S_y}$) is a property of unconfined aquifer. It has the dimensions of length and is usually expressed in metres. Large values of D indicate fast drainage. If D = ∞, the yield is instantaneous with the lowering of the water-table, i.e. the aquifer is unconfined without delayed yield.

Storativity Ratio (ω) This and the interporosity coefficient (λ) are the properties of fractured aquifers described in Sect. 7.2.2. Methods of estimating the hydraulic properties are described in Chap. 9.

8.2 Hydraulic Conductivity of Fractured Media

Fractures control the hydraulic characteristics of low permeability rocks, viz. crystalline, volcanic and carbonate rocks. Also in some clastic sedimentary formations, viz. sandstones, shales, glacial tills and clays, fractures form the main pathways for movement of fluids and contaminants.

In fractured rocks, a distinction can be made between hydraulic conductivity of fracture, K_f and of intergranular (matrix) material, K_m. As fractures form the main passage for the flow of water, the hydraulic conductivity of fractured rocks mainly depends on the fracture characteristics described in Chap. 2. The matrix permeability (k_m) in granite is estimated in the order of $10^{-19}\,m^2$ whereas fracture permeability (k_f) can vary from 10^{-12} to $10^{-15}\,m^2$ depending on the fracture aperture and interconnectivity. The role of some important parameters, e.g. aperture, spacing, stress, infilling (skin), connectivity etc. is discussed below (also see Sect. 19.6.3.2).

8.2.1 Relationship of Hydraulic Conductivity with Fracture Aperture and Spacing

The relationship between hydraulic conductivity (K_f) of a single plane fracture with aperture (a) is given by Eq. 8.15.

$$K_f = \frac{\gamma\, a^2}{12\mu} \tag{8.15}$$

The equivalent hydraulic conductivity of a rock mass, (K_s) with one parallel set of fractures is expressed by

$$K_s = \frac{a}{s} K_f + K_m = \frac{\gamma\, a^3}{12\, s\mu} + K_m \tag{8.16}$$

where s is fracture spacing. Usually K_m is very low except when rock matrix is porous and/or fractures are filled with impervious material. Therefore,

$$K_s = \frac{\gamma\, a^3}{12\, s\mu} = \frac{g\, a^3}{12\, \nu s} \tag{8.17}$$

where g is gravitational acceleration ($981\,cm\,sec^{-2}$) and ν is the coefficient of kinematic viscosity which is $1.0 \times 10^{-6}\,m^2\,s^{-1}$ for pure waer at 20°C.

In fractures with infillings, the hydraulic conductivity of fracture will depend on the permeability of the filling material, assuming that this permeability is still significantly greater than that of the rock matrix.

The hydraulic conductivity of a rock mass with three orthogonal joint sets with the similar spacing and constant aperture in all directions, in the three dimensional space, is given by Eq. 8.18 (Lee and Farmer 1993).

$$K_s = \frac{2\, \gamma\, a^3}{12\, s\mu} + K_m \tag{8.18}$$

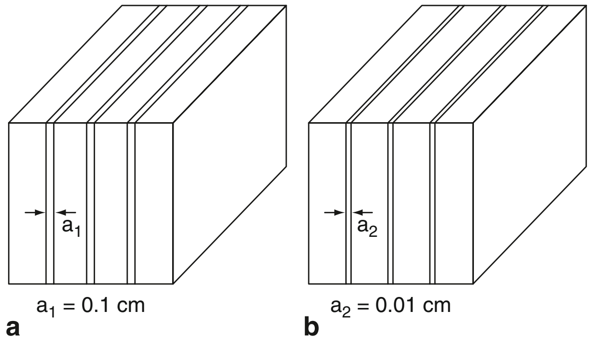

Fig. 8.4 Two sets of fractures with same spacing but different apertures

Figure 8.4 shows fractures with a common spacing of 1 joint per metre but with different apertures. In Fig. (8.4a) fracture apertures, $a_1 = 0.1\,cm$, and in Fig. (8.4b) $a_2 = 0.01\,cm$. Substituting these values in Eq. 8.17 will give equivalent hydraulic conductivity of K_1 to be $8.1 \times 10^{-4}\,m\,s^{-1}$ and K_2 for the second type will be $8.1 \times 10^{-7}\,m\,s^{-1}$ i.e. K_2 will be about three orders of magnitude less than that of set 1.

Figure 8.5 gives the hydraulic conductivity values of fractures with different apertures and frequencies. It could be seen that one fracture per metre with an aperture of 0.1 mm gives rock hydraulic conductivity of about $10^{-6}\,m\,s^{-1}$, which is comparable to that of porous sandstone. With a 1 mm aperture and the same spacing, the hydraulic conductivity will be $10^{-3}\,m\,s^{-1}$, similar to that of loose clean sand.

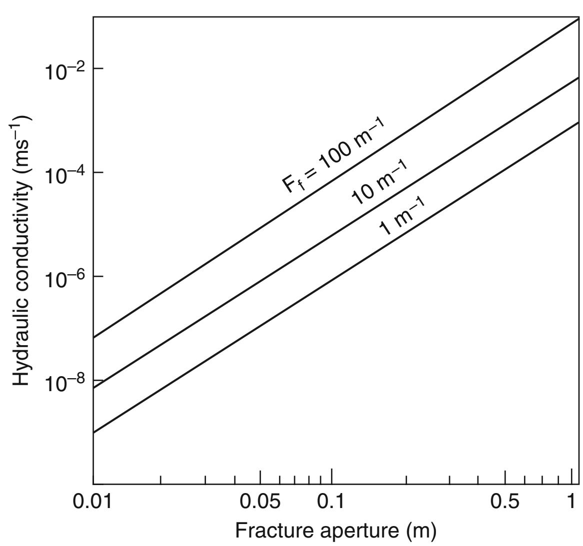

Fig. 8.5 Variation in hydraulic conductivity with fracture frequency and conducting aperture. (After Lee and Farmer 1993)

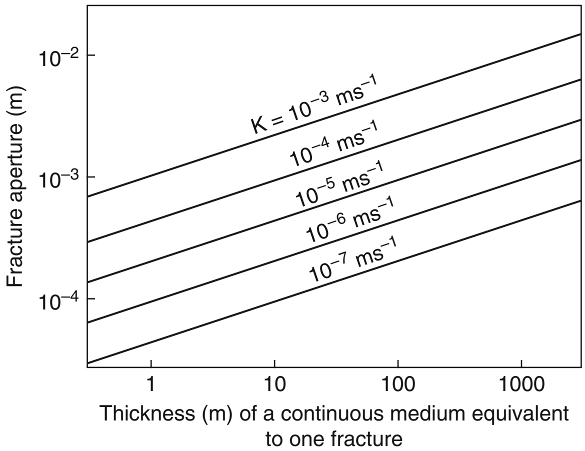

Fig. 8.6 Comparison between the hydraulic conductivity of the porous medium and that of the fractured medium as a function of aperture. (After Maini and Hocking 1977, reproduced with permission of the Geological Society of America, Boulder, Colorado, USA)

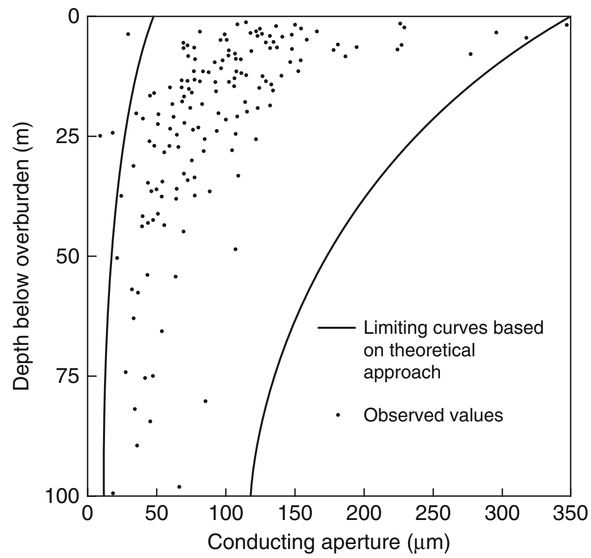

Fig. 8.7 Relationship between conducting aperture and depth based on field data. The right hand side envelope is based on initial conducting aperture of 350 μm and left hand side of 50 μm. JCS = 95.8 MN m^{-2}, JRC = 7.4. (After Lee and Farmer 1993)

The equivalence between hydraulic conductivity in a fractured rock and that of porous material is depicted in Fig. 8.6. As an example, the flow from a 10 m thick cross-section of a porous medium with a hydraulic conductivity of 10^{-4} m s^{-1} could be the same as from one single fracture with an aperture of about 1 mm. This demonstrates the large amount of flow which can be expected from fractures of even small openings.

A distinction is made between real or mechanical aperture (a_r) and conducting or hydraulic aperture (a_c). The real aperture is usually larger than conducting aperture if the fracture surfaces are rough. In smooth and wide fractures, the mechanical aperture and the conducting aperture will be equal. The empirical relation between a_r and a_c can be expressed as (Lee et al. 1996)

$$a_c = \frac{a_r^2}{JRC^{2.5}} \tag{8.19}$$

where a_r and a_c are in μm and JRC is Joint Roughness Coefficient, having a range from 20 (roughest surface) to 0 (smoothest surface). In natural fractures which are highly irregular, JRC may vary from 3 to12 from one part of the fracture to another. Increase in JRC results in an exponential decrease in flow rate.

Assuming a maximum initial conducting aperture (a_c) equal to 350 μm and minimum initial conducting aperture to be 50 μm and by considering other mechanical properties of similar rock types, Lee and Farmer (1993) obtained two curves showing variation in conducting aperture with depth (Fig. 8.7). These curves indicate a good agreement between the range of calculated and observed values. A similar trend of variation in fracture aperture was obtained by Oda et al. (1989).

8.2.2 Effect of Stress on Permeability

The mechanical behaviour and fluid flow in fractures is greatly influenced by the effective stress which is taken to be the normal stress on fracture minus the fluid pressure. Effective stress values are usually positive but in some cases as in hydrofracturing where fluid pressure exceeds the normal stress, effective stress values will be negative.

The combined effect of normal and shear stress on permeability and void structures is of importance in geotechnical and nuclear waste disposal studies and recovery of oil, gas and geothermal fluids from reservoirs (Rutqvist and Stephannson 2003). It is shown, both by theory and experimental studies, that stress reduces fracture aperture and thereby permeability of fractured rocks. Stress being a directional phenomenon, its state determines the relative permeability of different fracture sets in a rock mass. Fractures parallel to the maximum stress tend to be open, whereas those perpendicular to it tend to be closed.

Several experiments have been designed to estimate permeability variation in a variety of rocks under varying stress and thermo-mechanical conditions. (Brace 1978; Gale 1982a, 1982b; Oda et al. 1989; Read et al. 1989; Jouanna 1993; Azeemuddin et al. 1995; Indraratna and Ranjith 2001; Rutqvist and Stephannson 2003). Snow (Gale 1982a) proposed an empirical model of the form of

$$k = ko + \left(K_n a^2/s\right) \; (P - Po) \qquad (8.20)$$

where k is the permeability of horizontal fractures after loading, ko is the permeability at an initial pressure Po, K_n is the normal stiffness of the fracture; a and s are defined earlier.

Lee and Farmer (1993) quoting the work of Brace et al. showed a decrease in permeability of Westerly Granite from 350 nd at 10 MPa pressure to 4 nd at 400 MPa pressure (1 nd $=10^{-18}$ m^2). In Berea sandstone, a decrease in permeability from 10^{-10} m^2 to 10^{-11} m^2 due to increase in hydrostatic pressure from ambient to 30 MPa is reported by Read et al. (1989). The permeability reduction from an uniaxial strain test on Berea sandstone was estimated to be 20% but it was drastic (75%) in Indiana limestone which is attributed to pore collapse (Azeemuddin et al. 1995). Laboratory tests on shale, granite and sandstone show that shale has most stress-sensitive permeability while the sandstone is very sensitive at low stress but appears to attain a residual permeability at higher stress. The differences in the stress-permeability relationship in different rock types are explained by differences in pore shapes (Rutqvist and Stephansson 2003).

The effect of stress on permeability of jointed rocks also depends on the direction of stress in relation to joint orientation. Laboratory studies on jointed granite showed that an uniaxial stress of 12 MPa, parallel with the joint, raised fracture permeability (k_f), whereas 3 MPa, normal to the joint, decreased k_f to half of the initial value. However, even at the highest pressures (100 MPa), the permeability of the rock containing joints was at least a factor of 10^3 higher than the matrix permeability (k_m) (Brace 1978).

Studies indicate that when normal stress is applied to a natural fracture in the laboratory, there occurs a reduction in permeability indicating fracture deformation. The change in fracture permeability is higher in the initial stages (Fig. 8.8). Further investigations indicate that under confining stress, both in air and water, the average permeability decreases by almost 90% above 8 MPa compared to permeability values at zero confining pressure (Indraratna and Ranjith 2001). This is attributed to considerable reduction in joint aperture upto a certain value of confining stress. It is also noted that the reduction in permeability also depends on the roughness of fracture – the greater the roughness the lower the rate of reduction of permeability (Indraratna and Ranjith 2001). Varied responses of permeability during compression and decompression are also indicated (Van Golf-Racht 1982; NRC 1996). After a cycle of compression and decompression, the rock permeability may either return to the original permeability (Fig. 8.8), or may get reduced due to permanent deformation (Fig. 8.9). Laboratory experiments also show that in a mica schist, on application of normal stress (applied perpendicular to the foliation plane), the decrease in rate of flow is greater when normal stress increases than when the stress decreases (Fig. 8.10).

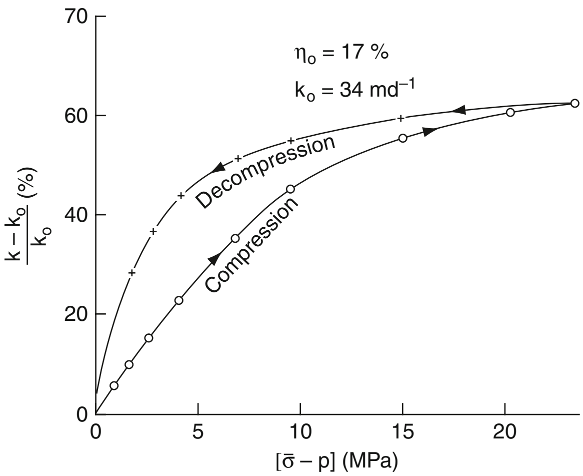

Fig. 8.8 Change in percent permeability with stress during compression and decompression in a fractured limestone. (After Van Golf-Racht 1982, with kind permission from Elsevier Science-NL, Amsterdam, the Netherlands)

Although effect of changes in normal stress on fracture permeability has been studied to a large extent, there have been very few controlled studies on the effect of shear stress on fracture permeability. The combined effect of normal and shear stresses on flow and void structures are not very well known so far (NRC. 1996; Rutqvist and Stephansson 2003). Such conditions are

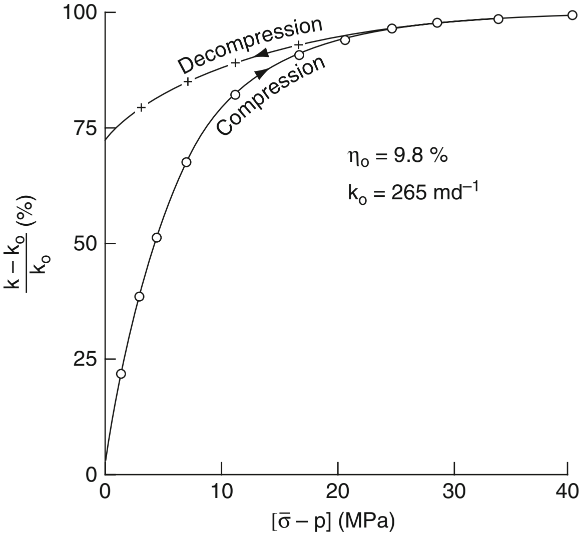

Fig. 8.9 Change in percent permeability with stress. Note that in this case the original permeability is not achieved after decompression. (After Van Golf-Racht 1982, with kind permission from Elsevier Science-NL, Amsterdam, The Netherlands)

likely to occur in civil and mining works such as underground excavations, dams and rock slopes.

It is expected that shear-stress will cause fracture dilation, especially in rough fractures due to displacement under low to moderate normal stress, resulting in significant changes in fracture permeability (Gale 1982a). In rocks like granites and quartzites, with porosity less than about 5%, the dilatancy effects have been found to be quite conspicuous. In granite, the permeability increased nearly fourfold, while in sandstone the increase was about 10–20%, but permeability of sand decreased considerably. The different behaviour of these materials indicates their varied response to stress (Brace 1978). Increase in permeability (k) and specific storage (S_S) due to the growth of dilatant cracks is also demonstrated by triaxial test (Read et al. 1989). On the other hand, in soft rocks, like mica schist, increase in shear stress showed a conspicuous decrease in rate of flow along the schistosity planes (Jouanna 1993) (Fig. 8.11).

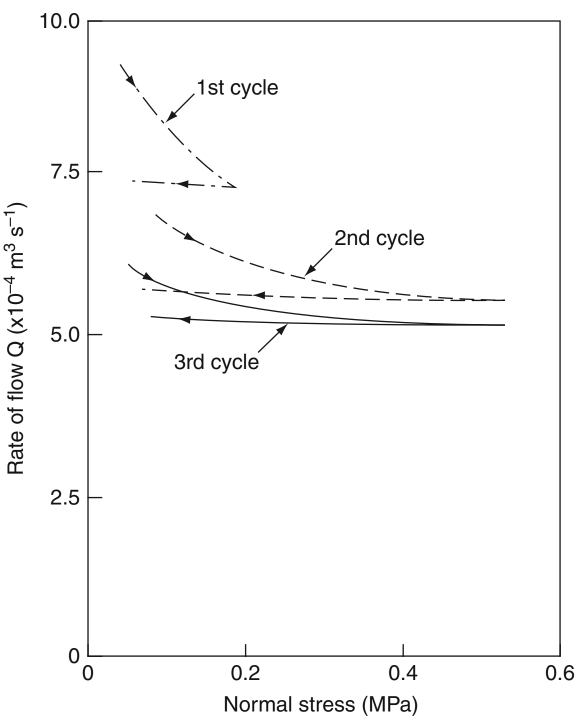

Fig. 8.10 Effect of normal stress on rate of flow in a mica schist. (After Jouanna 1993)

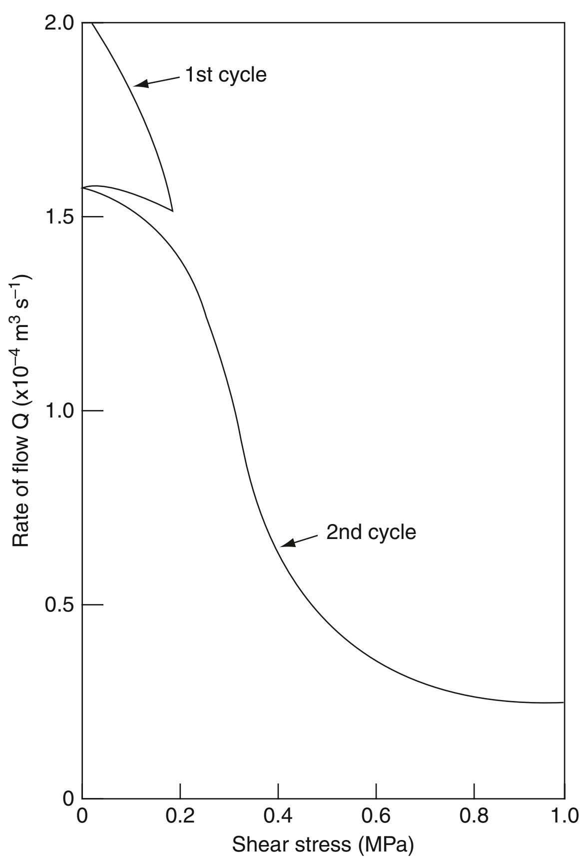

Fig. 8.11 Influence of shear stress on rate of flow in a mica schist. (After Jouanna 1993)

8.2.3 Relationship of Permeability with Depth

The depth dependence of permeability can be expressed by Eq. 8.21 given by Black (1987).

$$k = az^{-b} \tag{8.21}$$

where a and b are constants and z is the vertical depth below ground surface. Based on the data given by Snow (1968b) about the variation in the permeability of fractured crystalline rocks with depth from Rocky Mountain, USA, Carlsson and Olsson (1977) gave the Eq. 8.22.

$$K = 10^{-(1.6 \log z + 4)} \tag{8.22}$$

where K is the hydraulic conductivity in $\mathrm{m\,s^{-1}}$ and z is depth in metres. Similar empirical relations have been given by Louis (1974), and others. For example, based on tests in a number of wells in crystalline rocks in Sweden, Burgess (in Lee and Farmer 1993) gave Eq. 8.23.

$$\begin{aligned} \log K &= 5.57 + 0.352 \log Z \\ &\quad - 0.978 (\log Z)^2 + 0.167 (\log Z)^3 \end{aligned} \tag{8.23}$$

where K and Z have the same units as in Eq. 8.22.

Equation 8.23 can be transformed into Eq. 8.24 by relating stress and hydraulic conductivity on the basis of $\sigma = \gamma Z$

$$\begin{aligned} \log K &= 5.57 + 0.352 \left(\log \sigma/\gamma\right) \\ &\quad - 0.978 \left(\log \sigma/\gamma\right)^2 + 0.167 \left(\log \sigma/\gamma\right)^3 \end{aligned} \tag{8.24}$$

The decaease in permeability with depth in fractured rocks is usually attributed to reduction in fracture aperture and fracture spacing (Fig. 8.12). A least square fit to the packer-test data from boreholes in the granites of Stripa mine in Sweden, also indicated a general decreasing trend of permeability with increasing depth (Fig. 8.13a). The relationship between fracture frequeny and permeability from the same area is illustrated in Fig. 8.13b. A decrease in fracture aperture with depth is also reported from several other studies,

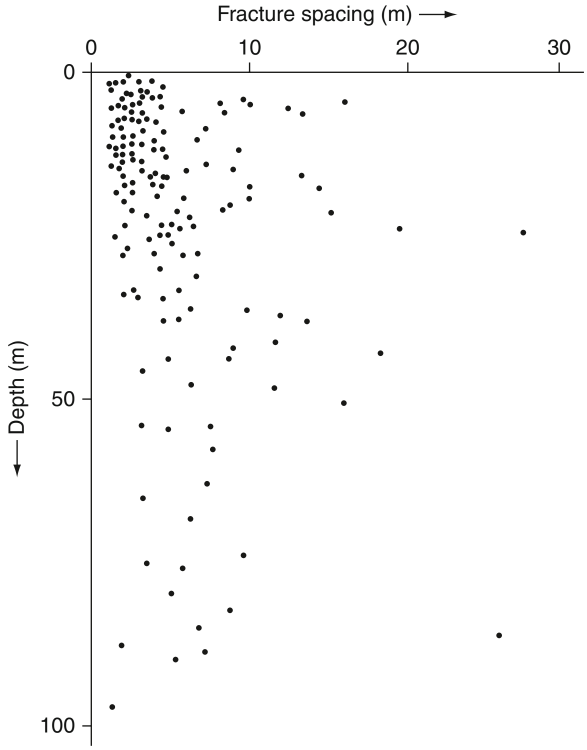

Fig. 8.12 Variation of fracture spacing with depth. (After Snow 1968a)

e.g. from a radioactive waste depository in andesite rocks in Taiwan (Lee et al. 1995).

Although, a decrease in permeability with increasing depth is demonstrated from several other places also, but this decrease may not be systematic, especially at greater depths (>50 m). The permeability can also vary by several orders of magnitude at the same depth (Fig. 8.14). Higher permeabilities at shallow depths (<50 m) can be attributed to greater influence of surficial phenomena like weathering etc. and development of sheeting joints due to unloading. Further, fractures at the same depth below the ground surface but with different orientations may be subjected to different stresses and therefore may have different permeabilities.

Even at depths of more than 1000 m, appreciable permeabilities are reported in fractured rocks. For examples, Fetter (1988) reported higher permeabilities from fractures at depths of 664–1669 m in granitic rocks of Illinois, USA. Recent studies under the Continental Deep Drilling Project in Germany (Kessels and Kuck 1995) and HDR experiments in the Rhine

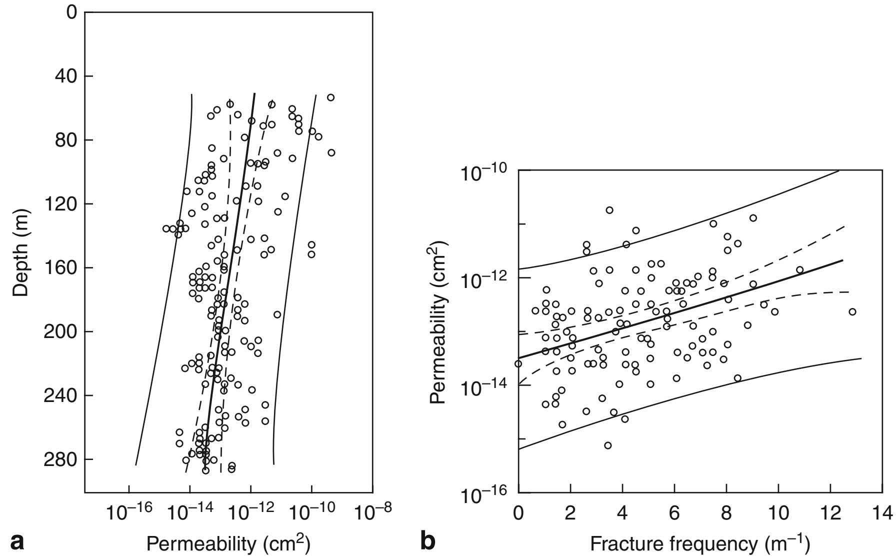

Fig. 8.13 Variation of permeability with: **a** depth, and **b** fracture frequency based on borehole packer tests in granites at Stripa, Sweden. (After Gale et al. 1982). The *central full line* in figures (a) and (b) is the least square regression line; on each side of this line the 95% confidence limits are shown for individual predicted values (*full lines*) and for mean predicted values (*dashed lines*)

Graben, France (Gerard et al. 1996; Stober and Bucher 2005) also show the existence of good hydraulic interconnection between adjacent boreholes through fractures even at a depth of more than 3000 m (also see Sect. 13.7.2).

On the basis of above discussion, it may be summarized that although, generally in fractured rocks a decrease in permeability with depth is observed at several places but there is not much justification of such an universal rule. Therefore, site specific studies are necessary.

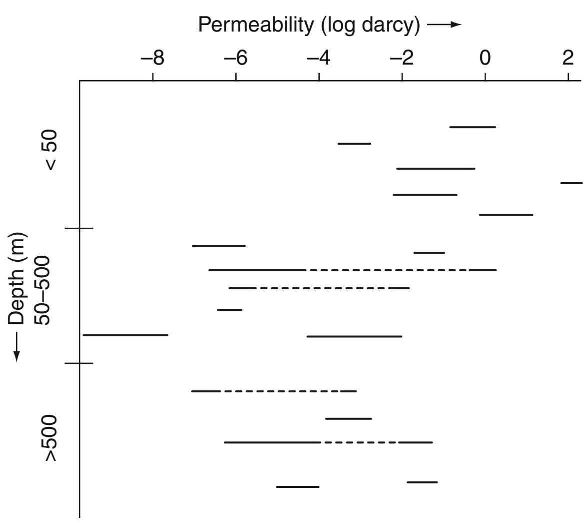

Fig. 8.14 Ranges of permeability with depth in crystalline rocks. (After Brace 1980)

8.2.4 Influence of Temperature on Permeability

Significant changes in rock temperature can occur due to natural weather conditions viz. alternate freezing and thawing and due to man induced changes. Formation of ice in fractures will cause extension of fractures and can also block the movement of water producing pressure build-ups. The influence of temperature on rock permeabilities is important for disposal of radioactive waste and in harnessing geothermal energy. An increase in temperature will cause a volumetric expansion of the rock material leading to reduction in fracture aperture and an overall decrease in rock permeability. Studies at Stripa mine in Sweden demonstrated a reduction in permeability of granites by a factor of three when temperature was increased by 25°C by circulating warm water. Similarly, in another experiment, a tenfold reduction in permeability was observed in a fractured gneiss when the temperature was increased by 74°C (Lee and Farmer 1993). On the other hand, thermal contraction causes development of new fractures during hydraulic fracturing in HDR experiments. However, doubts are created that these may not have significant positive long-term influence on development of geothermal energy from HDR systems (Zhao and Brown 1992) (see also Sect. 18.3). Changes in temperature may also change the effective stress in the rock mass. These stress changes may cause deformation of fractures as described earlier. Temperature changes will also cause precipitation and dissolution of minerals thereby affecting rock permeabilities.

8.2.5 *Effect of Fracture Skin on Permeability*

Fracture surfaces are usually covered with altered or detrital material, such as clay, iron or manganese oxides. These filling materials form fracture skin, which reduce the permeability of fractures and also influence the movement of solutes from the fractures into the matrix blocks. The presence of clay in the fractures may also increase the mechanical deformability of fractures.

8.2.6 *Interconnectivity of Fractures*

The interconnectivity of the different fracture sets is important for deciding the hydraulic continuity, which depends on fracture orientation, density, spacing and fracture size. Inter-connectivity can be expressed in terms of the ratio of average fracture spacing to fracture trace length. A ratio in the range of 1/20 to 1/50 has been suggested to indicate continuum (Lee and Farmer 1993). Long and Witherspoon (1985) studied numerically the effect of both the magnitude and nature of the fracture interconnection on permeability. The results showed that for a given fracture frequency, as fracture length increases, the degree of interconnection increases and thereby permeability also increases. Rouleau and Gale (in Lee and Farmer 1993) suggested an empirical interconnectivity index, I_{ij} between two fracture sets given by Eq. 8.25.

$$I_{ij} = \frac{l_i}{s_i} \sin \theta_{ij} \ (i \# j) \tag{8.25}$$

where l_i is the mean trace length for set i, s_i is the mean spacing of set i and θ_{ij} is the average angle between fractures of set i and j.

8.3 Anisotropy and Heterogeneity

Geological formations usually do not exhibit uniformity in their texture and structure either spatially or in different directions. Accordingly, their hydraulic characteristics such as hydraulic conductivity, and storativity also vary.

Anisotropy is usually a result of the rock fabric. In a porous rock consisting of spherical grains, the hydraulic conductivity will be the same in all directions and therefore it is said to be isotropic (Fig. 8.15a). On the other hand in a stratified formation, the constituent

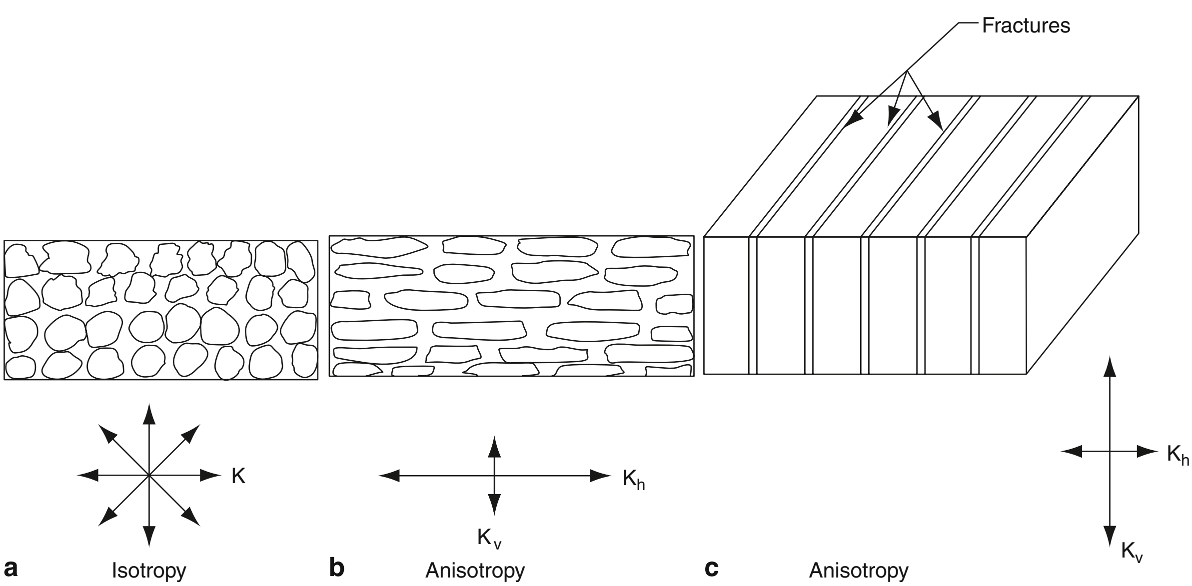

Fig. 8.15 Isotropic and anistropic aquifers: **a** isotropic sedimentary aquifer, **b** anisotropic sedimentary aquifer, and **c** anisotropic fractured aquifer

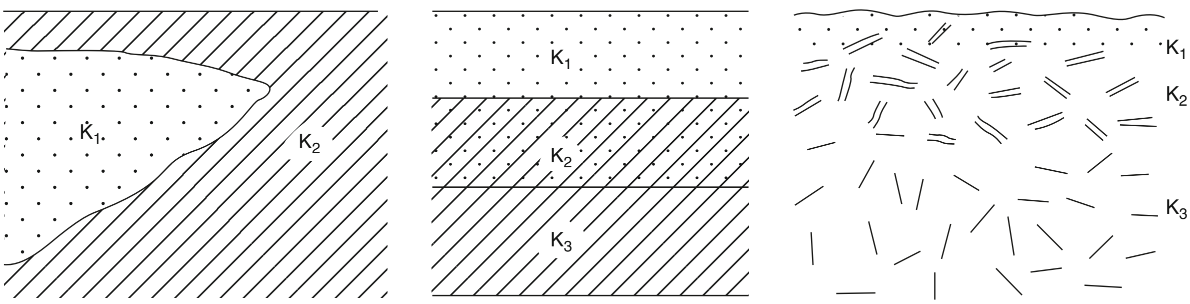

Fig. 8.16 Heterogeneous aquifers: **a** wedge-shaped aquifer **b** layered aquifer, and **c** fractured rock aquifer ($K_1 > K_2 > K_3$)

mineral grains of tabular nature, are usually laid with their longer axes parallel to each other. In such a case the hydraulic conductivity parallel to the bedding plane K_h, is higher than vetical conductivity, K_v (Fig. 8.15b). In fractured rocks, anisotropy is due to the presence of fractures which may have different orientations. The conductivity along the fracture, will be significantly higher than that normal to the fractures (Fig. 8.15c). In foliated rocks, like schists and phyllites also, conductivity parallel to foliation plane is greater than that normal to the foliation. Therefore, hydraulic conductivity is a tensorial property. The ratio of K_h to K_v may vary from 10^{-2} in fractured rocks to 10^3 in stratified sedimentary rocks as in the former permeability is mainly due to vertical fractures while in the latter the bedding planes which form easy passage for water are mostly horizontal. For analytical purpose, values of $K_h/K_v = 1$ to 10 are commonly used (Lee and Farmer 1993).

In a sedimentary rock sequence, it may be necessary to determine the average value of hydraulic conductivity normal to bedding (K_v) and parallel to bedding (K_h). If the total thickness of the sequence is H and the thickness of the individual layers are h_1, h_2, h_3, …, h_n, with corresponding values of the hydraulic conductivity K_1, K_2, K_3, …, K_n, then K_v and K_h can be obtained by

$$K_v = \frac{[H]}{h_1/K_1 + h_2/K_2 + h_3/K_3 + \cdots + h_n/K_n} \tag{8.26}$$

and

$$K_h = \frac{h_1K_1 + h_2K_2 + h_3K_3 + \cdots + h_nK_n}{[H]} \tag{8.27}$$

The heterogeneity in sedimentary formations could be due to lateral variation in aquifer thickness even if the hydraulic properties such as hydraulic conductivity and specific storage remain constant. Such situations are observed in wedge shaped aquifers (Fig. 8.16a), showing lateral variation in thickness, or where alternate beds of sediments with different textures are formed under varying depositional conditions. Figure 8.16b shows a layered sedimentary formation where individual bed has a homogeneous hydraulic conductivity K_1, K_2, … but the entire sequence shows a heterogeneous character due to vertical variation in hydraulic conductivity. In carbonate rocks, hetrogeneity can develop due to variation in degree of solution and in crystalline rocks it is a result of varying density of aperture sizes of fractures (Fig. 8.16c).

8.4 Representative Elementary Volume (REV)

REV is the smallest sample volume which is representative of the rock mass. The concept of REV is necessary to define the distribution of aquifer characteristics, such as hydraulic conductivity. In aquifer modelling also, it is necessary to have an idea of the minimum volume of rock which should be sampled having representative value of rock mass properties. In unfractured homogeneous rocks, the hydraulic conductivity tends to become constant beyond a particular rock volume. This least volume is known as REV (Fig. 8.17a). However, in fractured rock mass, hydraulic conductivity value may not become exactly constant with increase in sample volume but its variation may become rather insignificant (Fig. 8.17b). REV will also depend on

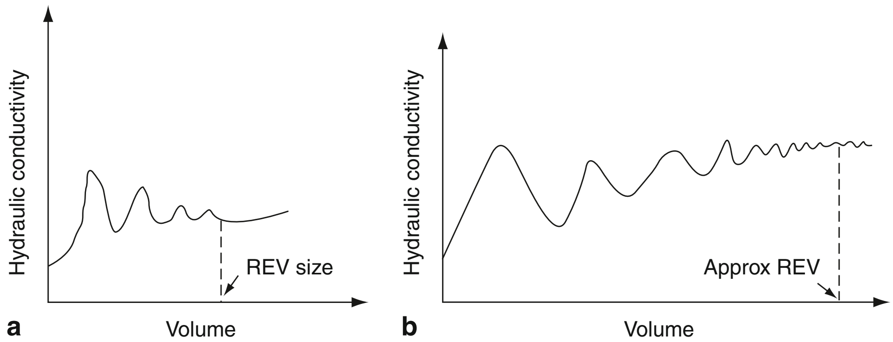

Fig. 8.17 Representative elementary volume (REV) in: **a** unfractured rock, and **b** fractured rock

the flow system and the geometrical parameters of the fractured medium.

In fractured rocks, the concept of REV will be applicable when there is a constant hydraulic gradient and linear flow lines as in a truly homogeneous anisotropic medium. Further, the following criteria must be met in order to replace a heterogeneous system of given dimensions with an equivalent homogeneous system for the purposes of analysis (Long et al. 1982).

(a) There is an insignificant change in the value of the equivalent permeability with small addition or substraction of the test volume.
(b) An equivalent permeability tensor exists which predicts the correct flux when the direction of gradient in a REV is changed.

Point (a) indicates that the REV size is a good representative of the sample considering rock mass heterogeneities and, point (b) implies that the boundary condition will produce a constant gradient throughout a truly homogeneous anisotropic sample.

REV increases in size with increase in discontinuity spacing. Therefore, in order to define REV, one should have sufficient knowledge of rock discontinuities. The influence of fracture geometry on REV is shown in Fig. 8.18. In granular rocks without discontinuities, small REV can be representative of the rock mass (Fig. 8.18a), but in fractured rocks, REV should be large enough to include sufficient fracture intersections to represent the flow domain (Fig. 8.18b). The size of the REV will be large compared to the size of the fractures lengths in order to provide a good statistical sample of the fracture population. However, in case of large scale features, such as faults and dykes, REV may not be feasible as it will be too large an area (Fig. 8.18c). This implies that the concept of REV may not be true and practical in every rock mass.

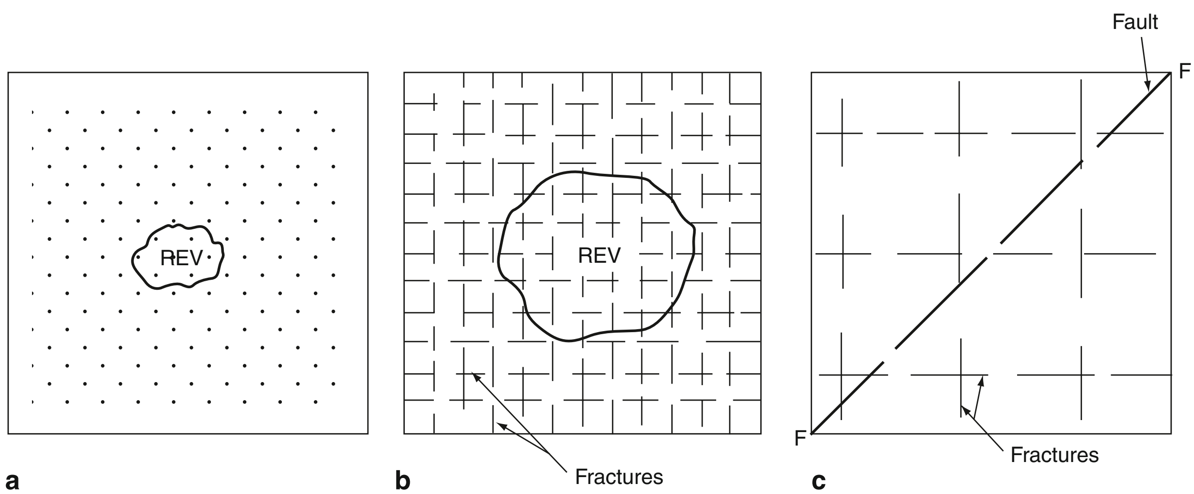

Fig. 8.18 REV in different rock conditions: **a** homogeneous porous rock, **b** fractured rock where REV includes sufficient fracture intersections, and **c** rocks with large scale discontinuities where REV is either very large or nonexistent

Summary
Important aquifer properties which control the occurrence and movement of groundwater are porosity (η), hydraulic conductivity (K), permeability (k), transmissivity (T) and storativity (S). In fractured rocks, a distinction is made between hydraulic properties of fractures and that of matrix, viz. K_f, K_m, S_f and S_m etc (the subscript f denotes properties of fractures and m that of matrix material).
The hydraulic conductivity of fractures depends on fracture characteristics, viz. aperture, spacing and interconnectivity. The fluid flow in fractures is also influenced by stress which in turn depends on the fracture orientation in relation to direction of stress. In general, permeability decreases with depth due to reduction in fracture aperture, though under certain specific tectonic conditions, high permeabilities are reported from very deep levels (>2 km).

Further Reading

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