Hydrogeology of Clastic Formations

16

Under clastic formations, we have included both unconsolidated and consolidated sediments. Unconsolidated sediments include various admixtures of boulder, sand, silt and clay deposits. These on consolidation form clastic sedimentary rocks, e.g. sandstone, siltstone and shale.

16.1 Unconsolidated Sediments

Most of the unconsolidated sediments were deposited during the last few million years, and are of Quaternary–Recent age. They have been formed under different sedimentary environments, viz. fluvial, glacial, eolian and marine. The coarse grained sediments, (gravel and sand) form potential aquifers due to their high hydraulic conductivity and storativity. Being incoherent in nature, unconsolidated or semi-consolidated sediments are largely unfractured, except some glacial deposits like clay tills.

16.1.1 Fluvial Deposits

The fluvial deposits are characterized by typical landforms, viz. alluvial fans, flood plain, and terraces etc. which can be identified on aerial photographs and space imageries. Fluvial deposits form ideal aquifers as they occur along river valleys and in areas of even topography with adequate recharge. Some of the river valleys are deep and narrow, e.g. the Jordan river valley along the border of Jordan and Israel, and the Owens river valley in California, USA. Others form broad plains, e.g. Indo-Gangetic basin in Indian subcontinent, and North Plain in China.

Buried valleys and palaeochannels are of special interest for groundwater development in hard rock terrains. They contain thick deposits of gravel and sand and are a result of fluvial and fluvio-glacial processes involving shifting of river courses due to either tectonic or climatic reasons. Palaeo-channels are identified by their typical sinuous serpentine shape and form. In an arid or semi-arid terrain, the palaeochannels may posses anomalous moisture and vegetation, in comparison to the adjoining areas. Table 16.1 gives a summary of physical conditions of palaceochannels and the corresponding spectral characters on remote sensing data.

An interesting example of palaeochannel is furnished by the '*lost*' Saraswati river in India, which is said to have been a mighty river in the olden *Vedic or pre-vedic* times (about 5000 years BC). It used to flow in the area now occupied by the Thar desert in western India. Here, the palaeochannels can be identified in most cases on the basis of higher moisture content and vegetation patterns (Fig. 16.1). Using Landsat images, the ancient course of the river Saraswati has been delineated for a distance of about 400 km (Yashpal et al. 1980; Baklliwal and Grover 1988). Some of the buried segments of river Saraswati are a potential source of groundwater (Venkateswarlu et al. 1990). In arid and hyper-arid regions, the buried channels are of added importance for water supply, owing to acute water scarcity conditions. The SIR-A data in Eastern Sahara has demonstrated the applicability of SLAR data for delineation of palaeochannels in hyper-arid regions (McCauley et al. 1982; Elachi et al. 1984). In the sand-covered limestone terrain of Rajasthan desert, India, Mehta et al. (1994) interpreted presence of relict rivers on the basis of ERS-SAR images (Fig. 16.2).

B. B. S. Singhal, R. P. Gupta, *Applied Hydrogeology of Fractured Rocks,*
DOI 10.1007/978-90-481-8799-7_16,

Table 16.1 Physical conditions of palaeochannels and their spectral characters on remote sensing data

Physical conditions associated with the palaeochannel	Spectral character
Higher surface moisture	Medium to dark in visible bands; dark on NIR and SWIR images; dark on SLAR images
Preferential denser vegetation cover	Dark on visible bands; light toned in NIR; light on SLAR images
Dry sand cover; no surface anomaly in terms of vegetation; bedrock occurs at a greater depth	Darker on longer wavelength SLAR images
Alluvial cover, thick gravelly river bed resulting in relatively less surface moisture and poor vegetation	Very light on visible and NIR bands

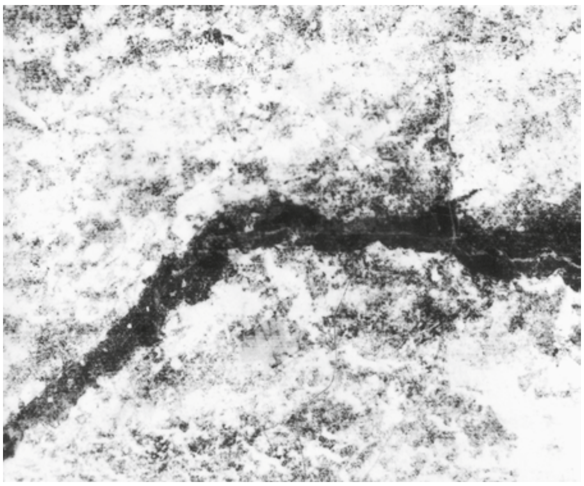

Fig. 16.1 Landsat MSS band-2 image showing the presence of palaceochannel of the 'lost' Saraswati river in the north-western India. The bed is about 6–8 km wide and is marked by dense vegetation

16.1.2 Glacial Deposits

Unconsolidated glacial deposits occur mainly in North America, northern Europe and Asia. Isolated occurrences are also observed on major mountain chains such as the Alps, the Andes, and the Himalayas, while close to the equator, glaciation was restricted to Mt. Kenya and Kilimanjaro in East Africa. Glacial deposits in North America and elsewhere have been studied in detail both as a source of water supply and also as host rock for the disposal of waste (Cherry 1989; Nilsson and Sidle 1996).

Glacial tills are unstratified and typically poorly sorted ice transported sediments in which the grain size varies from clay fraction to boulder size material. In some areas tills are thick and unfractured but wherever fractured, they provide active hydraulic connections and potential contaminant pathways (Hendry 1982; Ruland et al. 1991). Fracture spacing usually increases with depth thereby affecting vertical distribution of permeability (Fig. 16.3).

Fractures are often filled with calcite and gypsum. In some areas, thin rootlets are observed along the fractures to depths of 5–10 m below the ground surface (Freeze and Cherry 1979). Shallow fractures are regarded to be a result of alternate cycles of wetting and drying and freezing and thawing. The origin of deep open fractures in the unweathered clays is more problematic. These are attributed to stress changes caused by post-Pleistocene crustal rebound or to stress changes caused by advance and retreat of the last Pleis-

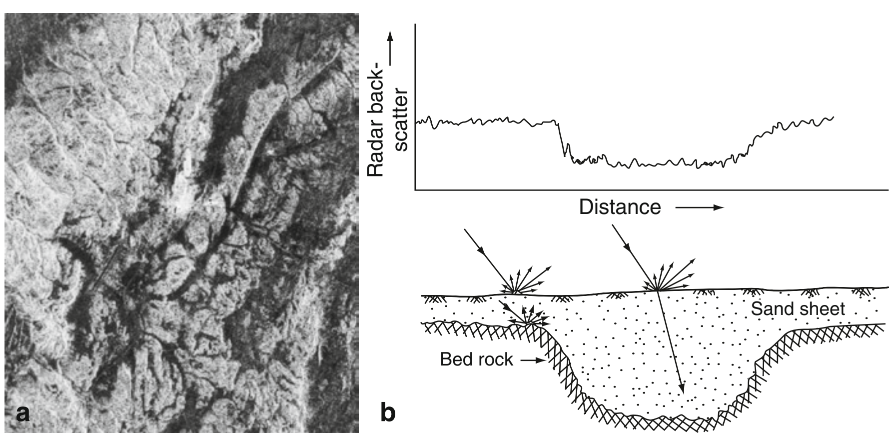

Fig. 16.2 **a** Palaeochannel as delineated on the ERS-SAR image of Rajasthan, Thar desert, India. The darker tones in the palaeochannel area are due to local thicker sand cover, as compared to the adjacent area (Courtesy: ESA). **b** Schematic interpretation. The lower radar backscatter is attributed to the thicker sand cover in the palaeochannel area

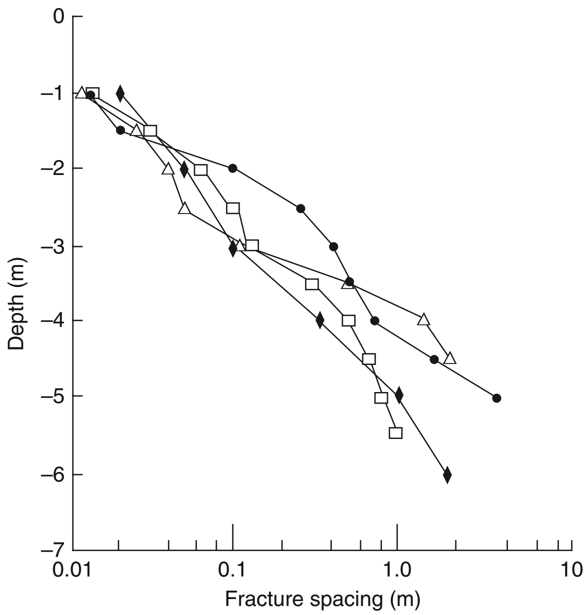

Fig. 16.3 Variation of fracture spacings in till with depth. (After Ruland et al. 1991)

tocene glacier (Cherry 1989). Volume changes due to geochemical processes, such as cation exchange, have been also suggested for the development of fractures (Freeze and Cherry 1979).

Glacial tills have porosities in the range of 25–45%; clay tills have higher porosities. Due to their unsorted character, the matrix (intergranular) permeabilities of glacial tills are low (10^{-10}–10^{-9} m s^{-1}). However, due to weathering and fracturing, they acquire higher hydraulic conductivities (K) of the order of 10^{-9}–10^{-6} m s^{-1} (Table 16.2). The upper parts of the till deposits usually have higher hydraulic conductivity due to greater intensity of fracturing and weathering (Ruland et al. 1991; Jones et al. 1992). The higher values of hydraulic conductivities in weathered and fractured tills will facilitate efficient drainage of agricultural lands but this will promote fast movement of pesticides and contaminants from landfills and other waste disposal pits. The anisotropy ratio (K_v/K_h) can have a wider range depending on the relative role of horizontal layering and vertical fracturing. Laboratory values of hydraulic conductivity measured from core samples is two to four order of magnitude smaller than that determined from field tests (Jones et al. 1992).

Detailed pumping tests with an array of observation wells, distributed radially, can provide better idea of aquifer parameters including anisotropy (Edwards and Jones 1993). Tracer injection tests are also carried out for the estimation of hydraulic parameters. Conservative tracers like chloride were used to estimate the potential of groundwater contamination in fractured tills in Denmark (Nilsson and Sidle 1996). Tritium profiles indicate active groundwater circulation in fractured tills upto a depth of 5–10 m (Ruland et al. 1991). The probability of monitoring the movement of contaminants in fractured clayey till can be enhanced by constructing large diameter horizontal or angled wells as vertical wells may not be able to intercept the fractures (Jorgensen et al. 2003).

Water in glacial aquifers is generally very hard and has high concentration of total dissolved solids and sulphate. In areas of active groundwater circulation, water quality is usually good. High concentration of SO_4 (1000–10 000 mg l^{-1}), reported from weathered clays, is attributed to oxidation of organic sulphur or pyrite (Cherry 1989). In such cases, the amount of SO_4^{2-} can be used as a natural tracer to study the downward solute migration through fractures in the unweathered clays.

The chemical evolution of groundwater in the fractures is influenced by the composition of the matrix and of pore water as well as the diffusion gradients. Studies in fractured glacial till indicate that different ions, e.g. Ca^{2+} and Cl^- behave differently due to

Table 16.2 Hydraulic properties of glacial deposits

Type of deposit	Location	T($m^2 d^{-1}$)	K(ms^{-1})	Sy	Source
Till (fractured clayey)	Manitoba and Montreal, Canada	–	10^{-10}–10^{-9}	–	Cherry (1989)
Till (fractured)	Alberta, Canada	–	5×10^{-9}–2×10^{-7}	–	Jones et al. (1992)
Till (weathered)	Iowa, USA	–	4×10^{-4} (av.)	0.04 (av.)	Edwards and Jones (1993)
Glacial outwash	Central Illinois, USA	43–1 66 488 (av.)	2×10^{-6}–7×10^{-6}	0.5	Walker and Walton (1961)
Glacio-fluvial	Michigan, USA	1440–1920	10^{-5}–10^{-4}	0.04–0.35	Kehew et al. (1996)

variation in their diffusion coefficients and adsorption by the matrix. Laboratory studies show that Ca^{2+} passes through the fractures more rapidly than Cl^- due to smaller diffusion coefficient of Ca^{2+} and adsorpition of Ca^{2+} within porous matrix (Grisak et al. 1980).

16.2 Consolidated Sediments

The common clastic sedimentary rocks are sandstone, siltstone and shale. These are formed in almost all environments including marine, fluvial, deltaic, lacustrine and eolian. Usually, sandstones and fine-grained clastic rocks, e.g. shales and siltstones, occur as alternating beds with varying thicknesses in most sedimentary sequences. Although sandstones being more permeable, are of main interest as a source of water supply, but their potentiality as well as water quality is greatly influenced by the composition of intergranular cementing material and properties of the interbedded shales and siltstones.

Sandstones being more resistant to erosion usually form hills, ridges and scarps while shales erode easily forming hill slopes and valleys. For the same reason, sandstones have low to medium drainage density but in shales drainage density is high. In sandstone, drainage pattern is rectangular or angular due to rock discontinuities. Shales show typical dendritic pattern. A comparison of landform and drainage characteristics in these two rock types is illustrated in Fig. 16.4.

16.2.1 Sandstones

The hydrogeological properties (porosity and permeability) of sandstones depend on their textural characteristics, which in turn is influenced by the depositional environments and subsequent changes due to cementation, consolidation and fracturing. The porosity, hydraulic conductivity and specific yield of sandstones is less than sands due to compaction and cementation. Common cementing materials are clays, carbonates, silica and iron oxides. As the degree of compaction depends on depth of burial, the porosity of sandstones decreases systematically with depth and age. Poorly cemented sandstones may have porosity of about 35%. Hydraulic conductivity of sandstones may vary from 1×10^{-6} to $1\times10^{-4}\,\mathrm{m\,s^{-1}}$ (Table 16.3). Geologically, older sandstones have lower porosity and hydraulic conductivity due to greater compaction and cementation, but there are several exceptions also. Stratification imparts anisotropy; hydraulic conductivity along the bedding plane is usually higher than across it. In Berea Sandstone, USA, permeability parallel to the bedding was found approximately four times higher than that normal to the bedding (Lee and Farmer 1993). At other places, fracturing may impart higher vertical conductivities.

Younger sandstones of Mesozoic and Tertiary ages form most productive aquifers covering large areas, e.g. Nubian sandstone in North Africa, Dakota sandstone in USA, Ranmark and Murray Group of sandstone in Australia, the Jurassic sandstone in Great Artesian Basin (GAB) in eastern Australia and Cud-

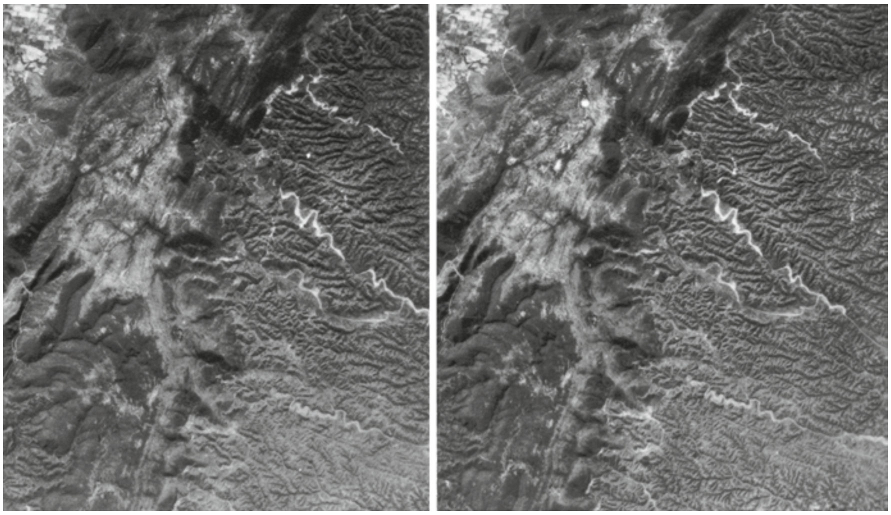

Fig. 16.4 Stereo aerial photographs showing sandstones and shales marked by differences in landform and drainage. (Courtesy A. White)

Table 16.3 Hydraulic properties of sandstones and argillaceous rocks

Rock type	Age	Location	$T(m^2d^{-1})$	S	$K(ms^{-1})$	Source
Charmuria sandstone (fractured)	Algonkian	Raipur, India	3–13	–	3×10^{-7}–3×10^{-6}	Khare (1981)
Berea sandstone	Mississipian	USA	–	–	3×10^{-6}	Read et al. (1989)
Shewood sandstone (fractured)	Triassic	UK	600 (av.)	0.05–0.25	–	Kimblin (1995)
Karoo sandstone (fractured)	Permo-Triassic	Botswana, South Africa	200	10^{-4}–10^{-2}	–	Bromley et al. (1994)
Nubian sandstone	Cretaceous	Egypt	100–15 000	3.5×10^{-4}	–	Shata (1982)
Renmark Group sandstone	Lower Tertiary	Victoria, Australia	800–900	1.8×10^{-4}	–	Lawrence (1975)
Cuddalore sandstone	Miocene	Tamilnadu, India	1000–5000	2×10^{-4}–5×10^{-4}	–	Gupta et al. (1989)
Gunderdehi shale (fractured)	Algonkian	Raipur, India	2–18	–	10^{-8}–10^{-7}	Khare (1981)
Shales and siltstones (fractured)	Upper Permian	Queensland, Australia	1–21 12 (av.)	–	–	Pearce (1982)
Siltstones (fractured)	Devonian	Appalachian Plateau, New York, USA	–	–	10^{-8}–10^{-6}	Merin (1992)
Brunswick shale (fractured)	Triassic	New Jersey, USA	708–933	10^{-4}	10^{-5}–10^{-4}	UNESCO (1972) Michalski (1990)
Pierre shale	Cretaceous	USA	–	–	10^{-14}–10^{-10}	Neuzil (1986)
Opaline clay	Mesozoic	Mt. Terri, Switzerland	–	–	10^{-11}	Neerdael et al. (1996)
Boom clay	Neogene	Belgium	–	$S_5=1.3\times10^{-5}m^{-1}$	10^{-12}	Put and Ortiz (1996)

dalore sandstone (Miocene–Pliocene age) in India. They form some of the best known confined aquifer systems extending over several thousand square kilometre and possess high transmissivities (Table 16.3). A case study of numerical modelling to estimate the effect of pumping of groundwater in Cuddalore sandstone (India) to facilitate open cast mining of overlying lignite is described in Sect. 19.5.5.

Groundwater in the Nubian sandstone is dated to be old (about 35 000 years) which was probably recharged during pluvial climate in Pleistocene time. Therefore, groundwater extraction in such areas should be planned carefully to avoid groundwater mining. Similar situation may exist in several other arid areas also.

Fractured Sandstones In firmly cemented sandstone, intergranular permeability is negligible but secondary permeability due to fracturing is of significance. Fracturing in sandstones could develop due to unloading, tectonic movements and at their contact with dolerite dykes. Fractures developed due to expansion, as a result of unloading, will be more prominent only upto a shallow depth (50 m or so), but tectonic fractures related with folding and faulting can be deep seated. Fracturing due to both tectonic movements and intrusion of dolerite dykes is reported from the Permo-Triassic sandstones of Karoo Supergroup in Botswana and other places in South Africa. As a result of fracturing, sandstones have acquired good transmissivity and specific yield (Table 16.3) (Bromley et al. 1994; Sami 1996). Triassic sandstones in UK also show much lower integranular hydraulic conductivity than the bulk aquifer conductivity as determined by pumping tests which is explained due to dominant fracture flow component (Hamill and Bell 1986). Similarly, other fractured sandstones also show moderate to high transmissivities (Table 16.3).

16.2.2 Shales and Siltstones

Fine-grained argillaceous clastic rocks, e.g. shales and siltstones are formed by compaction and lithification of clay and mud deposits. Porosity of freshly deposited clays and muds is high (50–80%), but due to compaction on burial, porosities are reduced to less than 30%. Shales usually have porosities in the range of 1.5–2.5%. Intergranular permeabilities in shales are low (10^{-13}–10^{-9} m s^{-1}) so that groundwater cannot move faster than few centimetre per century through intact shales (Table 16.3). Permeabilities of shales depend upon porosity, clay mineralogy, clay contents and texture. Permeabilities also depend upon fluid composition which affects clay swelling.

Even if the primary permeabilities of shales and siltstones are low, they are capable of transmitting large quantities of water and solutes over large contact areas by leakage across lithologic boundaries. For example, a 30 m thick siltstone bed with a hydraulic conductivity of 10^{-9} m s^{-1} having a hydraulic head difference of 3 m perpendicular to the bedding, will transmit about 3153 m^3 of water each year through each square kilometre of its surface areas.

Low permeability argillaceous rocks (clays and shales) have attracted greater attention of hydrogeologists in recent years from the point of view of disposal of high-level radioactive waste. With this in view, detailed investigations are in progress in the Underground Research Laboratories in the Boom Clay, Belgium and opaline clay/shale at Mt. Terri in Switzerland (Put and Ortiz 1996; Neerdael et al. 1996). Special *in situ* interference tests were designed in these underground laboratories to assess the hydraulic conductivity and storativity (Table 16.3).

Fractured Shales and Siltstones Fracturing can impart good hydraulic conductivity (10^{-7}–10^{-4} m s^{-1}) to shales, siltstones and other fine grained clastic rocks which are otherwise impervious (Table 16.3). In USA, notable examples include the Brunswick shale of Triassic age in New Jersey and adjoining states of New York and Pennsylvania, shales of Pennsylvanian age in the Central States and fractured siltstones of Devonian age in the Appalachain basin of New York state. It is opined that fractured shales represent a double porosity system, in which the fractures control the major amount of flow while the intervening porous blocks contribute slowly by transient flow (Neuzil 1986). Michalski and Britton (1997) have emphasised the importance of bedding-plane partings, particularly those enlarged by stress relief, to provide principal groundwater flow pathways and groundwater contamination in the Raritan unit of Newark Basin, USA. A similar conclusion is drawn from the Maquoketa aquitard (dolomitic shale of Ordovician age) in the southeastern Wisconsin where the hydraulic conductivity due to bedding plane fractures, extending to several kilometre, is about five orders of magnitude higher than the vertical permeability. Water quality also indicates the isolated nature of vertical fractures (Eaton et al. 2007). In a heterogeneous sedimentary sequence, thin incompetent beds, such as shales will be more intensely fractured as compared with thicker units of strong and resistant formations such as mudstones. This is evidenced from the Brunswick Formation, in USA, where mudstone horizons at shallow depths have lower values of K (10^{-6} m s^{-1}) while shales at deeper levels (20–45 m) are fractured forming aquifer horizons with hydraulic conductivity (K) varying from 2×10^{-5} to 5×10^{-4} m s^{-1} (Michalski 1990). Devonian siltstones of Appalachian plateau, USA exhibit two sets of fractures-subhorizontal bedding plane fractures and vertical fractures. Unlike vertical fractures, the spacing of bedding fractures increases with depth due to greater confining stress at deeper levels. Slug tests in these siltstones indicate that closer spacing of bedding plane fractures at shallow depths (<7 m) impart hydraulic conductivities (2×10^{-8}–2×10^{-6} m s^{-1}) which are 10–100 times greater than those obtained from deeper wells (>50 m) (Merin 1992).

16.2.3 Groundwater Development

Groundwater from clastic consolidated rocks is tapped both by dugwells and drilled wells, the latter being more common. The common method of water-well drilling in open textured sandstones is the hydraulic rotary method. In hard and well cemented sandstones, percussion and DTH methods are used (Chap. 17). Geologically younger sandstones are less cemented and therefore are more productive. The yield of 100–3000 m deep wells, in the Jurassic sandstones of the Great Artesian Basin in eastern Australia, is 0.05–0.1 $m^3 s^{-1}$. Here water from deeper wells has high temperature (1000 C). Well yields vary from less than $3\times10^{-4}\,m^3 s^{-1}$ in compact sandstones to $3\times10^{-2}\,m^3 s^{-1}$ in open textured rocks. Specific capacity of wells in Cambrian—Ordovician sandstones of northern Illinois, USA is in the range of 53–70 $m^2 d^{-1}$ indicating the hydraulic conductivity to be in the range of 1×10^{-5}–$1.8\times10^{-5}\,m\,s^{-1}$ (Walton and Csallany 1962). Well yields from fractured Charmuria sandstone of Precambrian age and Athgarh sandstone of Jurassic ages in India are of the order of $1.5\times10^{-3}\,m^3 s^{-1}$ and 10^{-3}–$10^{-2}\,m^3 s^{-1}$ respectively depending upon the degree of compaction and fracturing (Khare 1981; CGWB 1995b). In compact sandstones, well yields can be increased by shooting using explosives and to a less extent by acid treatment, viz. shooting increased the specific capacity of sandstone wells in northern Illinois by an average of 22–38% (Walton and Csallany 1962).

Integrated geophysical and geological studies have proved very valuable in the location of high yielding fracture zones in sandstone formation of Karoo supergroup especially in areas with complex geology obscured by overburden (Bromley et al. 1994). Figure 16.5 gives a comparison of the specific capacity of wells sited on fractures mapped by the above approach as compared with those in unfractured aquifer. This figure also indicates that major subvertical fractures identified by VLF survey have highest well yield. Studies by Sami (1996) in the Karoo sandstone of South Africa also show that wells located in the fracture zone in the vicinity of dolerite dykes and also those located along stream channels of higher orders have higher yields. Well yield in fractured shales will be low to moderate. However, in the absence of other good aquifers, wells in fractured shales may yield some water. Well yield from Lower Proterozoic fractured shales in Darwin Rural Area, NT, Australia is reported to be 1×10^{-3}–$3\times10^{-3}\,m^3 s^{-1}$.

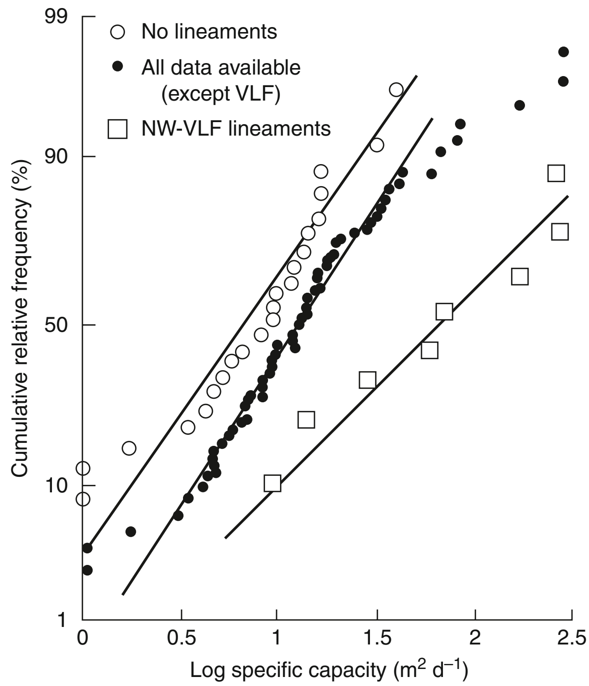

Fig. 16.5 Cumulative frequency of specific capacity of wells in fractured and non-fractured locations in Karoo sandstones, Botswana. Note that NW-VLF lineaments are most productive. (After Bromley et al. 1994, reprinted by permission of journal of Ground Water)

16.3 Water Quality

The chemical composition of groundwater in clastic rocks, as in other rock types, depends on their lithology and texture. In purely siliceous sandstones, the rocks consist almost wholly of quartz and therefore there are very few elements that can be dissolved in water. The groundwater in such rock types has low pH (about 6 or 5) and the HCO_3 content is also low; Ca, Mg, Cl and SO_4 are also in small concentrations; Na may exceed Ca. In rocks having lime or gypsum, the water has greater concentration of Ca and HCO_3. In arid and water logged regions, shallow groundwaters may have higher contents of salts due to evaporation (Greenman et al. 1967). In coastal areas, groundwater may show complex chemical characteristics due to mixing of fresh groundwater and sea-water as well as cation exchange. We return to this subject in Sect. 20.7.

Sandstone aquifers in large sedimentary basins show both horizontal (lateral) and vertical hydrochemical zoning (see Sect. 11.7.1). Groundwater in the recharge area is mainly of HCO_3 type which changes to SO_4-Cl type in the direction of flow. A decrease in SO_4 in the direction of flow is also reported from some places, e.g. groundwater in the Fox Hills sandstones in Dakotas, USA shows an increase in salinity but decrease in SO_4 from recharge to discharge areas. The source of SO_4^{2-} in recharge area is presumably due to pyrite and/or gypsum dissolution. The decrease in SO_4^{2-} appears to be a result of sulphate reduction in the presence of reduced carbon (lignite) while increase in chloride is attributed to cross flow of saline water from underlying Pierre shale (Thorstenson et al. 1979). Leakage from interbedded aquitards (clay, sillstone deposits) causing an increase in Cl^- content of groundwater is also reported from the Triassic Sherwood sandstone aquifer in northwest England (Kimblin 1995).

Shales and silstones being very fine and porous provide an enormous contact area with groundwater. The permeabilities are also very low, thereby increasing the residence time and greater opportunity for dissolution of rock material. Therefore, groundwater in these formations has high total dissolved solids ($>1000\,mg\,l^{-1}$) and the amount of SO_4^{2-} and Cl^- is more than HCO_3^- Both Ca–Mg and Na-rich waters are reported. In shales, SO_4^{2-} is the dominant anion while Cl^- is dominant in both shales and sandstones. Cation exchange is a characteristic phenomena in modifying the chemical characteristics of groundwater due to the presence of clay minerals. Clay and shales also play an important role in acting as semi-permeable membrane (osmotic filters) inspite of their low permeability as described in Sect. 11.6. Such a process is believed to be responsible for the occurrence of saline waters in non-marine sediments (Back and Hanshaw 1965; Neuzil 1986).

Movement of tracers and contaminants in fractured sandstone and shale will behave like in other dual porosity aquifers with rapid movement in fractures and diffusion into low permeability but porous matrix. Fracture skins formed due to the chemical alteration of sandstones into oxides of iron and manganese considerably lower the matrix rate of diffusion. These aspects have been discussed in Sect. 7.4.1.

Summary

Clastic formations include porous sedimentary rocks (sandstones, shales etc.) which have undergone consolidation and cementation to varying degree. The consolidated and cemented clastic formations have low porosities and permeabilities. Therefore their hydrogeological characteristics are similar to those of crystalline rocks.

Glacial deposits viz. tillites have been studied in detail both as a source of water and also as host rocks for the disposal of waste. Sandstones, siltstones and shales are characterised by bedding planes and are also fractured exhibiting varying degree of primary/secondary porosities and permeabilities. Younger sandstones of Mesozoic and Tertiary ages generally form more productive aquifers. Older sandstones of Precambrian age are often less productive. Shales and siltstones have low permeability, but fracturing can impart greater hydraulic conductivity. The water quality depends on lithology and permeability. Groundwater in sandstones is usually of good quality but in shales and siltstones, TDS is high on account of large residence time of water. Hydrochemical zoning is reported from large sedimentary basins.

Further Reading

Davis SN, DeWiest RJM (1966) Hydrogeology. John Wiley & Sons, Inc., New York.

Kresic N (2007) Hydrogeology and Groundwater Modeling. 2nd ed., CRC Press, Boca Raton, FL.

Walton WC (1970) Groundwater Resource Evaluation. McGraw-Hill, New York.