2 Groundwater Investigations

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2.1 Introduction

Successful drainage depends largely on a proper diagnosis of the causes of the excess water. For this diagnosis, one must consider: climate, topography, pedology, surface water hydrology, irrigation, and groundwater hydrology (or hydrogeology). Each of these factors – either separately, or more often in combination – may create a surface drainage problem (flooding, ponding) or a subsurface drainage problem (shallow watertable, waterlogging), or a combination of these problems. Most of these factors are treated separately elsewhere in this book.

In this chapter, we shall concentrate on some hydrogeological aspects of drainage problems, particularly those of subsurface drainage. Although each area has its own specific groundwater conditions, a close relationship exists between an area's groundwater conditions and its geological history. So, first, we shall discuss this relationship. For more information reference is made to Davis and De Wiest 1966 and Freeze and Cherry 1979.

We shall then explain how to conduct a groundwater investigation, describing how to collect, process, and evaluate the groundwater data that are pertinent to subsurface drainage, for more details see UN 1967. For further reading Mandel and Shiftan 1981, Matthess 1982, Nielsen 1991, Price 1985 and Todd 1980 are recommended.

2.2 Land Forms

Land forms are the most common features encountered by anyone engaged in drainage investigations. If land forms are properly interpreted, they can shed light upon an area's geological history and its groundwater conditions.

The two major land forms on earth are mountains and plains. Plains are areas of low relief and have our main interest because they usually have rich agricultural resources that can be developed, provided their water management problems are solved. This does not mean that the drainage engineer can neglect the mountains bordering many such plains. Mountain ranges are the source of the sediments occurring in the plains. They are also the source of the rivers that carry the detritus to the plains where it is deposited when the rivers flood.

Not all plains are of the same type; their source area, transporting agent, and depositional environment will differ. We recognize the following types of plains:
- Alluvial plains, formed by rivers;
- Coastal plains, formed by the emergence of the sea floor;
- Lake plains, formed by the emergence of a lake floor;
- Glacial plains, formed by glacial ice.

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The above list is not complete but covers the types of plains that are important for agriculture; they are also areas where drainage problems are common. Their main geological features and groundwater conditions will now be briefly described. For more information on land forms reference is made to Thornbury 1969.

2.2.1 Alluvial Plains

Alluvial plains are formed by rivers and usually have a horizontal structure. The larger alluvial plains, however, are downwarped and often faulted. They may contain alluvial sediments that are hundreds or even thousands of metres thick. Along the course of the river, from the mountains to the sea, we recognize three types of alluvial plains:
- Alluvial fan;
- River plain;
- Delta plain.

Alluvial Fans
An alluvial fan is a cone-like body of alluvial materials laid down by a river debouching from a mountain range into lowland (Figure 2.1). Alluvial fans are found in both the humid and the arid zones. They occur in all sizes, the size being largely determined by the size and geology of the river catchment and the flow regime of the river. Boulders, gravel, and very coarse sand dominate the sediments at the fan head. The sediments usually become finer towards the distal end of the fan, where they may be silt or fine sandy loam.

Because of the variability of a river's flow regime, alluvial fans are subject to rapid changes, with channels shifting laterally over a wide area, and channels alternating, cutting, and filling themselves. The mud-flow deposits extend as a continuous sheet over large areas, whereas the sand and gravel deposits are usually restricted to former channels.

Because of the very coarse materials at the head of the fan and the many diverging stream channels, substantial quantities of river water percolate to the underground. The head of the fan is therefore a recharge zone. The middle part of the fan is mainly a transmission zone. The distal end of the fan is a zone of groundwater discharge. Owing to the presence of mud-flow deposits of low permeability, the groundwater in the deep sand and gravel deposits is confined (i.e. the water level in a well that penetrates the sand and gravel layers stands above these layers).

The watertable, which is deep in the head of the fan, gradually becomes shallower towards the distal end, where it may be at, or close to, the surface. In arid zones, substantial quantities of groundwater are lost here by capillary rise and evaporation, which may turn the distal fan into a true salt desert.

Whereas subsurface drainage problems are restricted to the distal fan, surface drainage problems may occasionally affect major parts of an alluvial fan, especially when heavy rains evacuate highly sediment-charged masses of water from the mountains.

River Plains
River plains are usually highly productive agricultural areas. Rivers occur in different

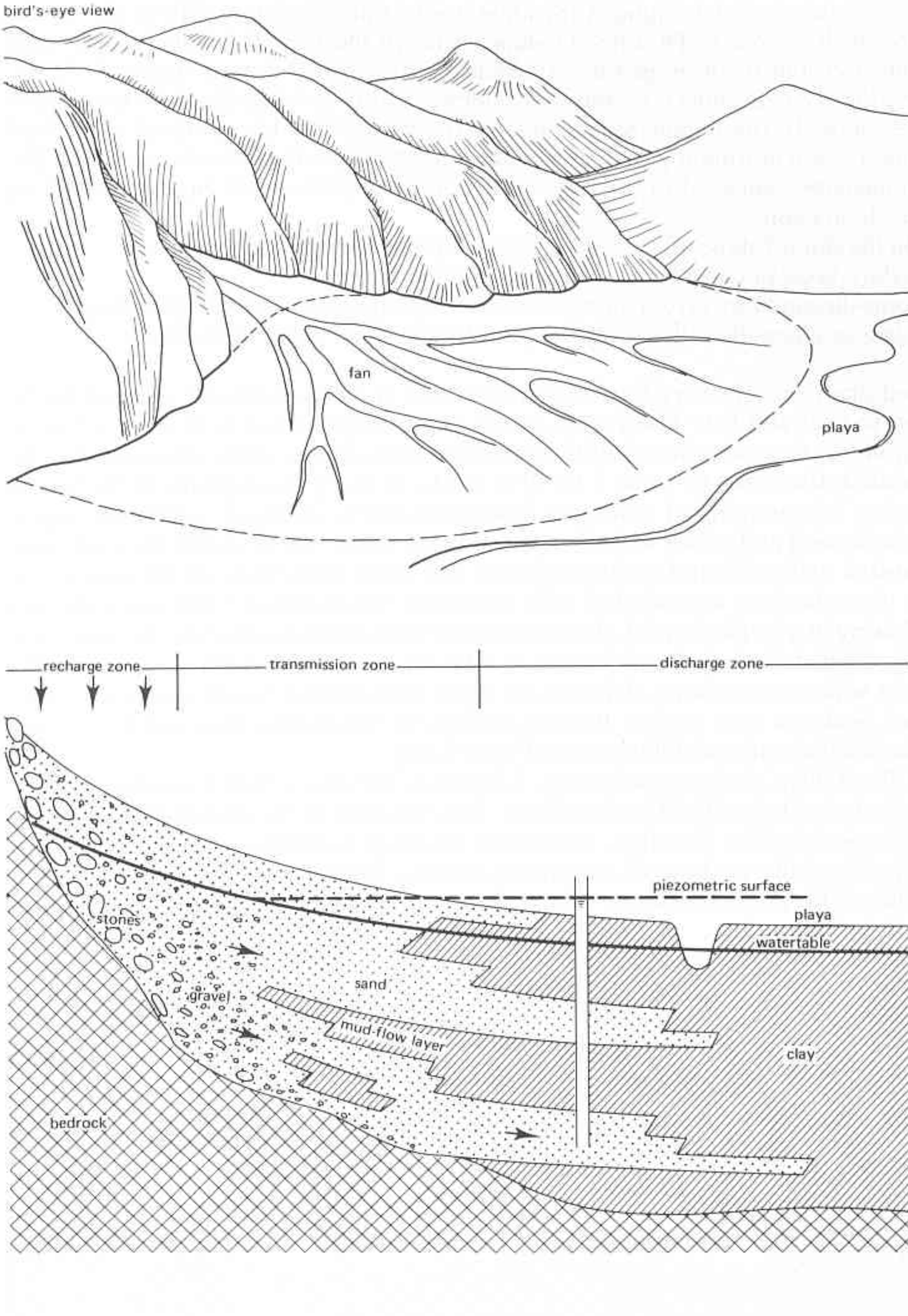

Figure 2.1 Bird's-eye view and idealized cross-section of an alluvial fan showing the flow of groundwater from the recharge zone to the discharge zone

sizes and stages of development (Strahler 1965), but a common type is the graded river, which has reached a state of balance between the average supply of rock waste to the river and the average rate at which the river can transport the load. At this stage, the river meanders, cutting only sideways into its banks, thus forming a flat valley floor. In the humid zones, most of the large rivers have formed wide flood plains. Typical morphological features of a flood plain are the following (Figure 2.2):

- Immediately adjacent to the river are the natural levees or the highest ground on the flood plain;
- On the slip-off slope of a meander bend are point bars of sandy materials;
- Oxbow lakes or cut-off meander bends contain water;
- Some distance away from the river are backswamps or basin-like depressions;
- Terraces along the valley walls represent former flood plains of the river.

Flood plain deposits vary from coarse sand and gravel immediately adjacent to the river, to peat and very heavy clays in the backswamps or basins (Figure 2.3A). In the quiet-water environments of these basins, layers of peat, peaty clay, and clay are deposited; they may be some 5 to 10 m thick. In many flood plains in the humid climates, a coarsening of sediments downward can be observed; continuous layers of coarse sand and gravel underlie most of those plains. These coarse materials were deposited under climatological conditions that differ from those of the present. At that time, the river was supplied with more rock waste than it could carry. Instead of flowing in a single channel, the river divided into numerous threads that coalesced and redivided. Such a river is known as a braided river. The channels shift laterally over a wide area; existing channels are filled with predominantly coarse sand and gravel, and new ones are cut. In cross-section, the sediments of braided rivers show a characteristic cut-and-fill structure (Figure 2.3B).

A flood plain, as the name implies, is regularly flooded at high river stages, unless it is protected by artificial levees (dikes). Deforestation in the catchment area of the river aggravates the floodings. Subsurface drainage problems are common in such plains, especially in those of the humid climates. Shallow watertables and marshy conditions prevail in the poorly drained backswamps and other depressions. Seepage

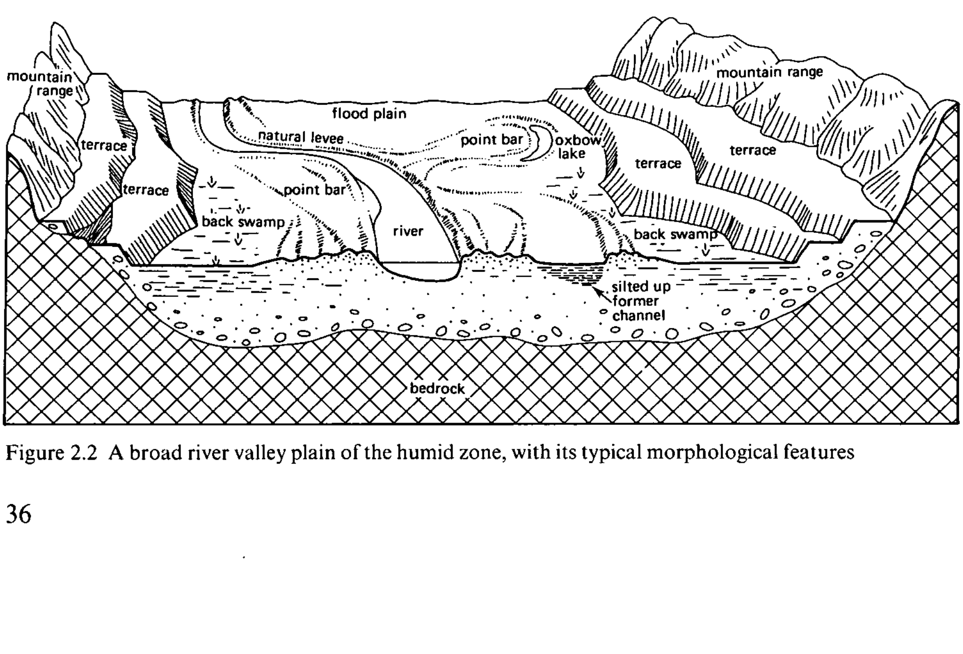

Figure 2.2 A broad river valley plain of the humid zone, with its typical morphological features

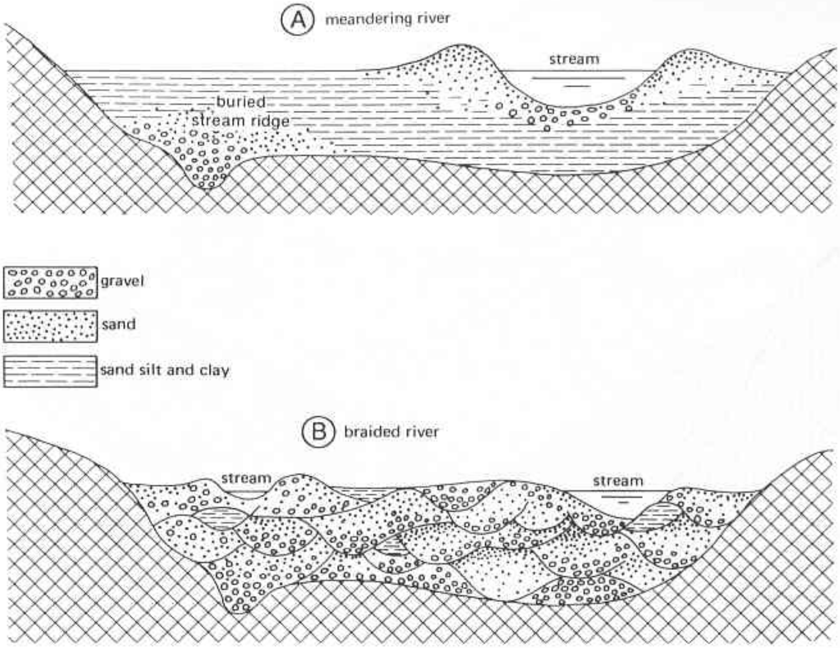

Figure 2.3 Cross-sections over a valley with:
A: A meandering river; B: A braided river

from the river, when it is at high stage, contributes to the subsurface drainage problems of the plain. (This will be discussed further in Chapter 9.)

Delta Plains

Deltas are discrete shoreline protuberances formed where rivers enter the sea or a lake and where sediments are supplied more rapidly than they can be redistributed by indigenous processes. At the river mouth, sediment-laden fluvial currents suddenly expand and decelerate on entering the standing water body. As a result, the sediment load is dispersed and deposited, with coarse-grained bedload sediment tending to accumulate near the river mouth, whilst the finer-grained sediment is transported offshore in suspension, to be deposited in deeper water.

Delta plains are extensive lowlands with active and abandoned distributary channels. Between the channels is a varied assemblage of bays, flood plains, tidal flats, marshes, swamps, and salinas (Figure 2.4).

In the humid tropics, a luxuriant vegetation of saline mangroves usually covers large parts of a delta plain. In contrast, delta plains in arid and semi-arid climates tend to be devoid of vegetation; salinas with gypsum ($CaSO_4$) and halite (NaCl) are common, as are aeolian dune fields.

Some delta plains are fluvial-dominated because they are enclosed by beach ridges at the seaward side; others are tide-dominated because tidal currents enter the distributary channels at high tides, spilling over the channel banks and inundating the adjacent areas. But even in areas with moderate to high tidal ranges, the upper

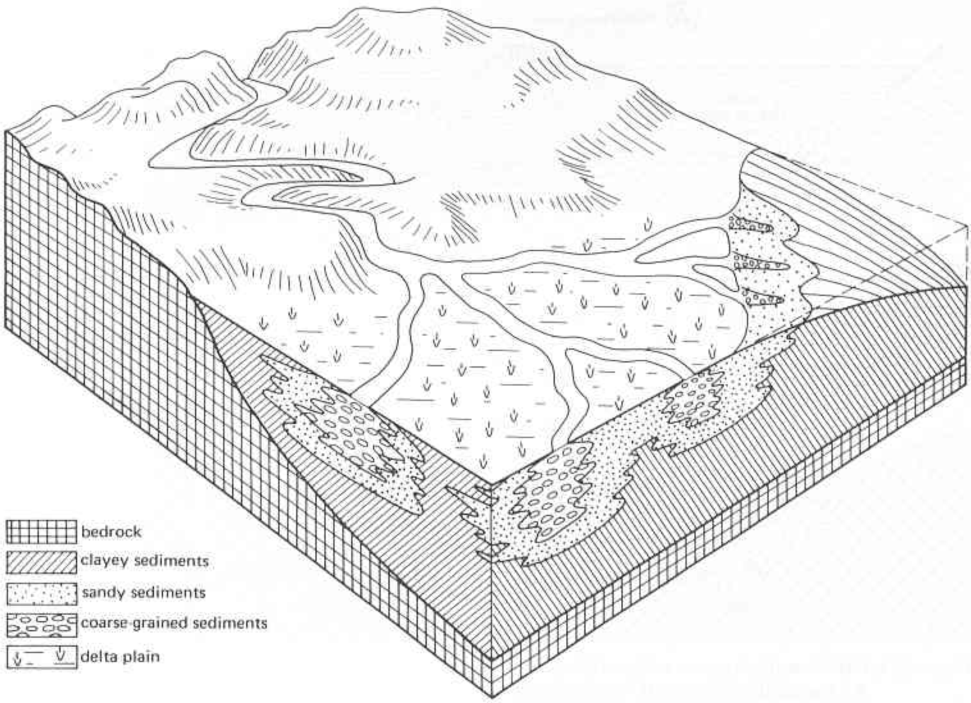

Figure 2.4 A typical delta

delta plain is fluvial-dominated; the lower delta plain may be tide-dominated, and the delta front may be either tide- and/or wave-dominated.

The groundwater in the upper delta is usually fresh and the watertable is relatively deep; problems of subsurface drainage and soil salinization do not occur there. In the lower delta, however, such problems do occur because the watertable is close to the surface, the groundwater is salty or brackish, and salinized soils are widespread, especially in arid deltas.

Figure 2.5 shows the groundwater flow and groundwater conditions in a delta plain.

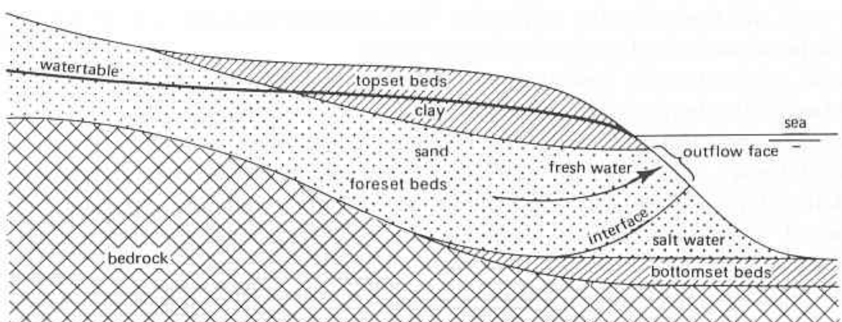

Figure 2.5 Cross-section of a delta plain, showing the interface of fresh and salt water and the outflow face at the delta front

The fresh groundwater of the upper delta moves seaward because its phreatic level is above sea level; it flows out in a narrow zone at the delta front. Owing to diffusion and dispersion, the initially sharp interface between fresh and salt water bodies gradually passes into a brackish transition layer. The rate at which this transition layer develops depends on various factors, one of which is the permeability of the delta sediments (Jones 1970).

2.2.2 Coastal Plains

From a geological point of view, coastal plains are an entirely different type of plain because they are recently-emerged parts of the sea floor. If the sea floor emerged in the remote geological past, coastal plains can be found far from the sea and are then called 'interior plains'. The structure of coastal plains can be simple, consisting of a continuous sequence of beds, or complex as a result of several advances and retreats of the sea (Figure 2.6).

Coastal plains may be narrow or even fragmentary strips of the former sea floor,

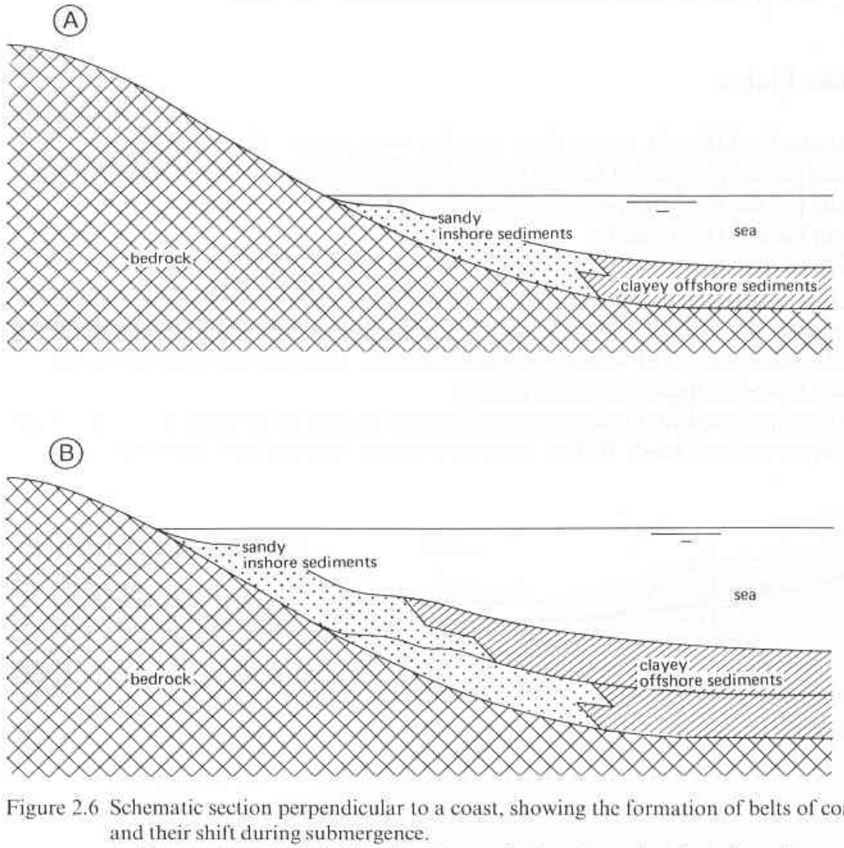

Figure 2.6 Schematic section perpendicular to a coast, showing the formation of belts of coastal sediment and their shift during submergence.
A: Constant sea level; B: Sea rises to new level and remains there, depositing sediments as in A. New sediment overlies the older sediment (after Longwell et al. 1969)

exposed along the margins of an old land area, or they may be vast, almost featureless plains, fringing hundreds of kilometres of coastline. Gently sloping coastal plains are attacked by the waves offshore and, as a result, sand bars may form parallel to the coast. Some plains are enclosed by dunes at the seaward side. The land enclosed by the dunes and by the natural levees of rivers that traverse the coastal plain is a true basin, containing lakes and swamps. In contrast to the upper coastal plain, the lower coastal plain may suffer from severe surface drainage problems.

The groundwater conditions of coastal plains are complex. The watertable in the upper part of the plain is usually deep, but gradually becomes shallower towards the coast. The soils in the upper part are usually sandy and permeable, so that subsurface drainage problems do not occur. Here we find outcrops of the sand layers that dip seaward under the lower coastal plain (Figure 2.7).

Because there is a seaward fining of sediments and because the sand layers are (partly) overlain and underlain by clayey deposits, the groundwater in the sand layers of the lower plain is confined. The seaward thinning of the sand layers contributes to the confined groundwater conditions. As a result, there is upward seepage from the deeper sandy layers through the clay layers to the surface. Both surface and subsurface drainage problems are common in lower coastal plains.

2.2.3 Lake Plains

Lakes originate in different ways; they may be river-made, glacial, volcanic, fault-basin, and landslide lakes. Most lakes are small and ephemeral. Those formed in active tectonic areas persist for long periods of geological time. Some lakes fall dry because of a change in their water balance.

Emerged lake floors are flat, almost featureless plains. In arid zones, surface water collects in the lower parts of the plains, where it evaporates, leaving behind the suspended sediments, mixed with fine salt crystals (halite, gypsum, carbonates). The sediments of such ephemeral water bodies commonly build up clay-surface plains of extraordinary flatness; these are called playas.

Most of the clastic sediment deposited in lakes is transported there by rivers, either in suspension or as bed load. Where the river water spreads out upon entering the

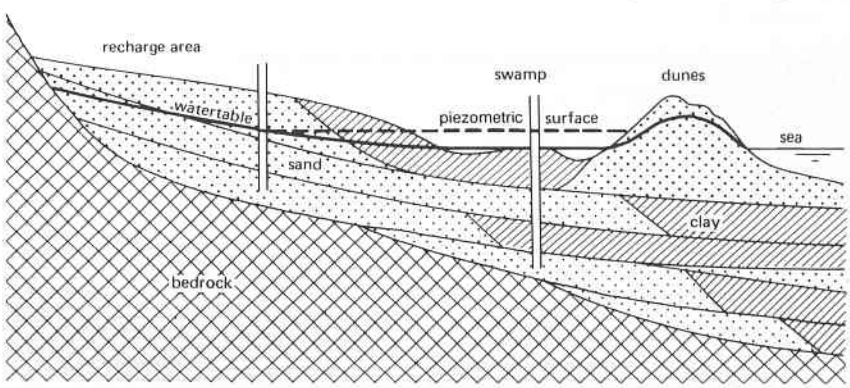

Figure 2.7 Groundwater conditions of a coastal plain

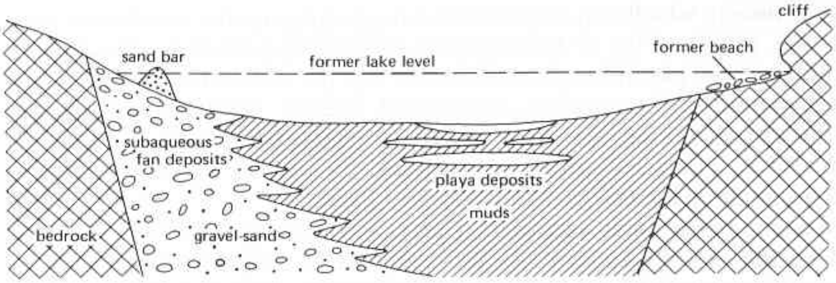

Figure 2.8 Geological section through the sediments of a former lake

lake, bed load, and coarse materials of the suspended load will be deposited first. Further away from the river mouth, there will be a distal fining of sediment, as in marine deltas. Where water densities allow underflow of the river water, coarse sediment may be spread out over the lake floor. Eventually, this may lead to the development of subaqueous fans in front of the river mouth. In the deepest parts of lake basins, clastic deposition is almost entirely from suspension.

Several morphological features can provide evidence of the former existence of a lake (Figure 2.8): for instance, a cliff on the hard rocks bordering the former lake, lake terraces at the foot of the cliff, and beaches and sand ridges, representing spits and sand bars formed by wave action near the lake margin.

Lake plains in humid climates usually have a shallow watertable that must be controlled if the land is to be used for agriculture. In lake plains in arid climates, the watertable is deep, except where rivers enter the former lake. If the river water is used for irrigation, the watertable may be very shallow in most of the irrigated areas. Because of the low permeability of the lake sediments, groundwater movement is slow, resulting in large watertable gradients at the margins between irrigated and non-irrigated lands.

2.2.4 Glacial Plains

At present, some 10 per cent of the earth's surface is covered by glacial ice. During the Quaternary glaciation, the maximum coverage was about 30 per cent and the resulting sediments were distributed over vast areas.

A characteristic sediment of the basal or subglacial zone – the contact zone of ice and bed(rock) – is till, a glacially-deposited mixture of gravel, sand, and clay-sized particles. Its texture is extremely variable. Massive tills occur as sheets, tongues, and wedges, but locally cross-sections appear as blankets (Figure 2.9). The erodibility of the substrata largely controls the thickness of till deposits. Where there was considerable local relief, the till is thick in depressions and in deeply incised channels, and is thin or absent on highs.

Another characteristic subglacial deposit is that of eskers. These are formed in melt-water tunnels at the base of the glacial ice. Eskers are mainly composed of sandy materials and appear in the landscape as ridges.

Around the margin of glacial ice – and strongly influenced by the ice – is the proglacial environment, which may be glaciofluvial, glaciolacustrine, or glaciomarine environments (i.e. produced by, or belonging to, rivers, lakes, or seas, respectively). Typical examples of the glaciofluvial environment are the outwash deposits, which may cover substantial areas marginal to glaciated regions. Outwash deposits are composed of sand and gravel; they form where meltwater is abundant and coarse material is available. These sediments are deposited by braided streams. Extensive outwash fans occur along the margin of a stationary glacier.

Lakes are a common feature in a proglacial landscape. They develop because rivers are dammed by the ice and because of the formation of irregular topography by glacially-deposited or eroded land forms. In cross-section, filled glacial lakes show the classical structure of steeply inclined foreset beds, chiefly made up of coarse sands, and gently sloping bottomset beds made up of fine-grained lake-floor deposits (mud, silt, and clay). The lake-bottom deposits typically consist of varves, each varve

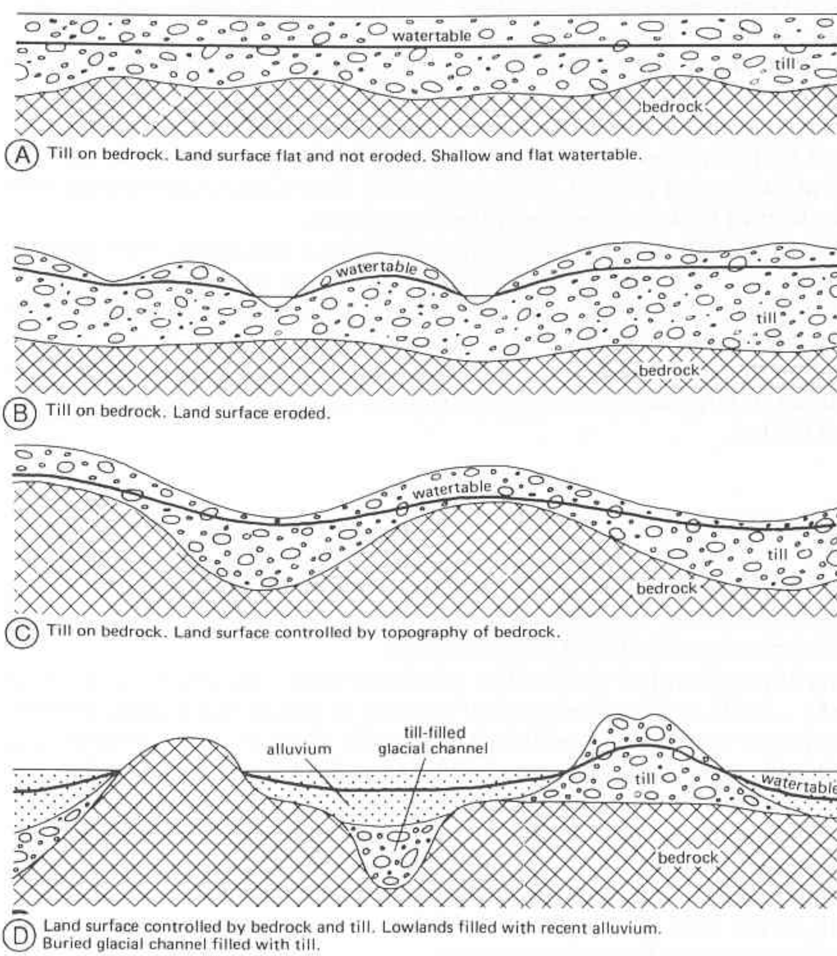

Figure 2.9 Cross-sections showing various relationships between bedrock and till deposits

consisting of a lower coarse layer of very fine sand-to-silt and an upper layer of clay.

Both climatic instability and powerful winds contribute to aeolian activity around a glacier or ice sheet. Sand reworked from glacial deposits is shaped into aeolian dunes. Silt is also readily picked up from glacial outwash and is deposited down-wind as loess, which is generally well-sorted and poorly stratified or non-stratified. Loess deposits can form blankets up to tens of metres thick and can extend hundreds of kilometres from the ice margin.

Young glacial plains are often characterized by poor drainage. This is mainly due to the low permeability of the till deposits and the undeveloped drainage systems. Once a drainage system has developed, the drainage of the higher grounds will be better. Excess water from rainfall and from surface runoff may collect in local depressions and lowlands, causing flooding and high watertables.

2.3 Definitions

2.3.1 Basic Concepts

The water in the zone of saturation, called groundwater, occurs under different conditions.

Where groundwater only partly fills an aquifer (a permeable layer), the upper surface of the saturated zone, known as the watertable, is free to rise and fall. The groundwater in such a layer is said to be unconfined, or to be under phreatic or watertable conditions (Figure 2.10A).

Where groundwater completely fills a permeable layer that is overlain and underlain by aquicludes (i.e. impermeable layers), the upper surface of the saturated zone is fixed. Groundwater in such a layer is said to be confined, or to be under confined or artesian conditions (Figure 2.10B). The water level in a well or borehole that penetrates into the permeable layer stands above the top of that layer, or, if the artesian pressure is high, even above the land surface. Relative to a chosen reference level,

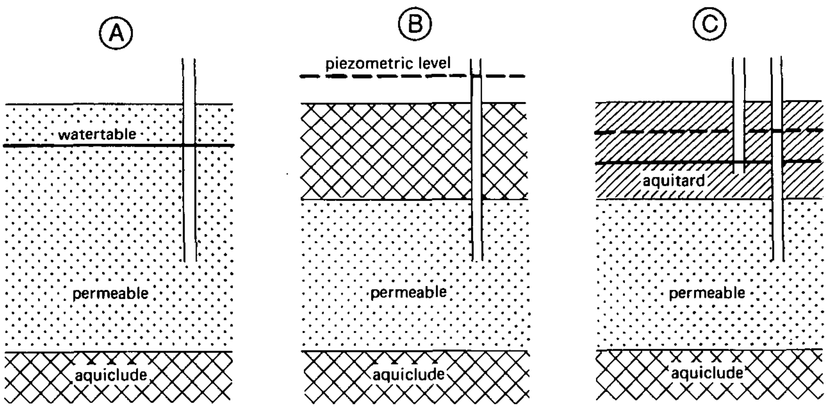

Figure 2.10 Different groundwater conditions:
A: Unconfined (watertable, phreatic) conditions; B: Confined (artesian) conditions; C: Semi-confined conditions

the height of the water column in the well is called hydraulic head, being the sum of pressure head and elevation head. (This will be discussed further in Chapter 7.) Truly impermeable layers are not common in nature; most fine-textured layers possess a certain, though low, permeability.

Where groundwater completely fills a permeable layer that is overlain by an aquitard (a poorly permeable layer) and underlain by an aquiclude or aquitard, the groundwater in the permeable layer is said to be semi-confined (Figure 2.10C). In the overlying aquitard, the groundwater is under unconfined conditions because it is free to rise and fall. The water level in a well or borehole that penetrates the permeable layer stands above the top of that layer or even above the land surface if the pressure is high. This type of groundwater condition is very common in nature. Three different situations can be recognized:

- The water level in the well stands at the same height as the watertable in the overlying aquitard, which means that there is no exchange of water between the two layers;
- The water level stands above the watertable in the aquitard, which means that there is an exchange of water between the two layers, with the permeable layer losing water to the aquitard; here, one speaks of upward seepage through the aquitard;
- The watertable stands below the watertable in the aquitard; here, the permeable layer receives water from the aquitard, and one speaks of downward seepage (or natural drainage) through the aquitard.

These three groundwater conditions can also occur in combination (e.g. in stratified soils, where permeable and less permeable layers alternate, or, besides seepage through the poorly permeable top layer, there may also be seepage through an underlying layer).

In some areas, drainage problems can be caused by a perched watertable. The watertable may be relatively deep, but a hardpan or other impeding layer in the soil profile creates a local watertable above that layer or hardpan.

It will be clear by now that an area's groundwater conditions are closely related to its geological history. As the geology of an area is usually variable, so too are its groundwater conditions. In a practical sense, the problem is one of identifying and evaluating the significance of boundaries that separate layers of different permeability. In subsurface drainage studies, we are interested primarily in the spatial distribution and continuity of permeability.

In solutions to groundwater problems, permeable, poorly permeable, or impermeable layers are usually assumed to be infinite in extent. Obviously, no such layers exist in nature; all water-transmitting and confining layers terminate somewhere and have boundaries. Hydraulically, we recognize four types of boundaries:

- Impermeable boundaries;
- Head-controlled boundaries;
- Flow-controlled boundaries;
- Free-surface boundaries.

An impermeable boundary is one through which no flow occurs or, in a practical sense, a boundary through which the flow is so small that it can be neglected. A permeable sand layer may pass laterally (by interfingering) into a low-permeable or impermeable clay layer, as we saw in Section 2.2.2 when discussing the geology of coastal plains. As the transition of the two layers, we have an impermeable boundary. Other examples of impermeable boundaries are a sand layer that terminates against a valley wall of impermeable hardrock, or a fault that brings a permeable and an impermeable layer in juxtaposition (Figure 2.11).

A head-controlled boundary is a boundary with a known potential or hydraulic head, which may or may not be a function of time. Examples are streams, canals, lakes, or the sea, which are in direct hydraulic contact with the water-transmitting layers. Note that an ephemeral stream in arid zones is not a head-controlled boundary because part or most of the time it does not contain water, and when it does, the water in the stream is not in contact with the groundwater, which may be many metres below the stream bed.

A flow-controlled boundary, also called a recharge boundary, is a boundary through which a certain volume of groundwater enters the water-transmitting layer per unit of time from adjacent strata whose hydraulic head and permeability are not known. The quantity of water transferred in this way usually has to be estimated from rainfall and runoff data. A typical example is the underflow entering the head of an alluvial fan (Section 2.2.1). Note that an impermeable boundary is a special type of flow-controlled boundary: the flow is zero.

A free-surface boundary is the boundary between the zones of aeration and saturation (i.e. the watertable is free to rise and fall). In subsurface drainage studies, this boundary is of primary importance.

Many subsurface drainage projects cover only a portion of an alluvial, coastal, or glacial plain. Project boundaries therefore do not usually coincide with hydraulic boundaries.

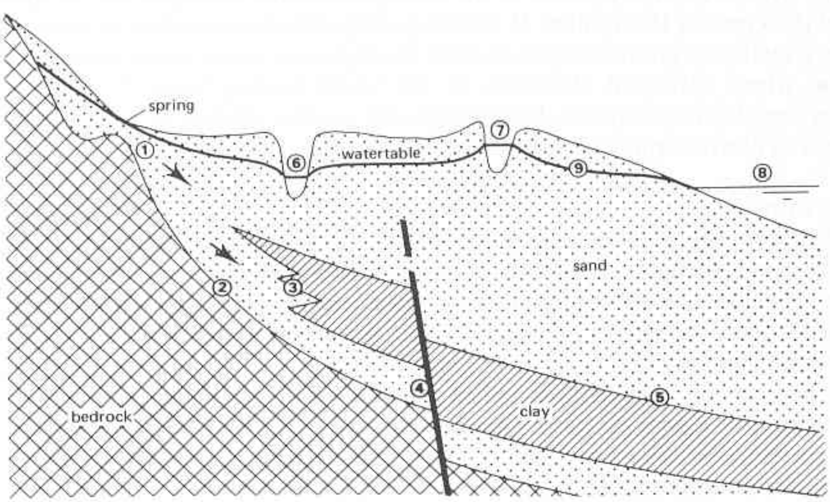

Figure 2.11 Examples of hydraulic boundaries:
1: Flow-controlled boundary; 2, 3, 4, and 5: Impermeable boundaries; 6, 7, and 8: Head-controlled boundaries; 9: Free-surface boundary

2.3.2 Physical Properties

To describe the flow of water through the different layers, one needs data on the following physical properties: hydraulic conductivity, saturated thickness, transmissivity, drainable pore space, storativity, specific storage, hydraulic resistance, and the leakage factor. These will be briefly explained.

Hydraulic Conductivity

The hydraulic conductivity, K, is the constant of proportionality in Darcy's law and is defined as the volume of water that will move through a porous medium in unit time under a unit hydraulic gradient through a unit area measured at right angles to the direction of flow. Hydraulic conductivity can have any units of length/time (e.g. m/d). Its order of magnitude depends on the texture of the soil (Chapter 3) and is affected by the density and viscosity of the groundwater (Chapter 7).

Saturated Thickness

For confined aquifers, the saturated thickness, H, is equal to the physical thickness of the aquifer between the aquicludes above and below it (Figure 2.10B). The same is true for a semi-confined aquifer bounded by an aquiclude and an aquitard (Figure 2.10C). In both these cases, the saturated thickness is a constant. Its order of magnitude can range from several metres to hundreds or even thousands of metres. For unconfined aquifers (Figure 2.10A), the saturated thickness, D', is equal to the difference in level between the watertable and the aquiclude. Because the watertable is free to rise and fall, the saturated thickness of an unconfined aquifer is not constant, but variable. It may range from a few metres to some tens of metres.

Transmissivity

The transmissivity, KH, is the product of the average hydraulic conductivity, K, and the saturated thickness of the aquifer, H. Consequently, the transmissivity is the rate of flow under a hydraulic gradient equal to unity through a cross-section of unit width and over the whole saturated thickness of the water-bearing layer. It has the dimensions of length^2/time and can, for example, be expressed in m^2/d. Its order of magnitude can be derived from those of K and H.

Drainable Pore Space

The drainable pore space, μ, is the volume of water that an unconfined aquifer releases from storage per unit surface area of aquifer per unit decline of the watertable. Small pores do not contribute to the drainable pore space because the retention forces in them are greater than the weight of water. Hence, no groundwater will be released from small pores by gravity drainage.

Drainable pore space is sometimes called specific yield, drainable porosity, or effective porosity. It is a dimensionless quantity, normally expressed as a percentage. Its value ranges from less than 5 per cent for clayey materials to 35 per cent for coarse sands and gravelly sands (Chapter 3).

Storativity

The storativity, S, of a saturated confined aquifer is the volume of water released