from storage per unit surface area of aquifer per unit decline in the component of hydraulic head normal to that surface. In a vertical column of unit area extending through the confined aquifer, the storativity, S, equals the volume of water released from the aquifer when the piezometric surface drops over a unit decline distance. The storativity is a dimensionless quantity. It is the algebraic product of an aquifer thickness and specific storage and its value in confined aquifers ranges from 5×10^{-5} to 5×10^{-3}.

Specific Storage
The specific storage, S_s, of a saturated confined aquifer is the volume of water that a unit volume of the aquifer releases from storage under a unit decline in head. This release of water under conditions of decreasing hydraulic head stems from two mechanisms:
- The compaction of the aquifer due to increasing effective stress;
- The expansion of water due to decreasing water pressure (see also Chapter 9).

For a certain location, the specific storage can be regarded as a constant. It has the dimension of length^{-1}.

Hydraulic Resistance
The hydraulic resistance, c, characterizes the resistance of an aquitard to vertical flow, either upward or downward. It is the ratio of the saturated thickness of the aquitard, D′, and its hydraulic conductivity for vertical flow (K) and is thus defined as

$$c = \frac{D'}{K} \tag{2.1}$$

The dimension of hydraulic resistance is time; it can, for example, be expressed in days. Its order of magnitude may range from a few days to thousands of days. Aquitards with c-values of 1000 days or more are regarded as aquicludes, although, theoretically, an aquiclude has an infinitely high c-value.

Leakage Factor
The leakage factor, L, describes the spatial distribution of leakage through an aquitard into a semi-confined aquifer, or vice versa. It is defined as

$$L = \sqrt{KHc} \tag{2.2}$$

High values of L originate from a high transmissivity of the aquifer and/or a high hydraulic resistance of the aquitard. In both cases, the contribution of leakage will be small and the area over which leakage takes place, large. The leakage factor has the dimension of length and can, for example, be expressed in metres.

2.4 Collection of Groundwater Data

To obtain data on the depth and configuration of the watertable, the direction of groundwater movement, and the location of recharge and discharge areas, a network of observation wells and/or piezometers has to be established.

2.4.1 Existing Wells

Existing wells offer ready-made sites for watertable observations. Many villages and farms have shallow, hand-dug wells that can offer excellent observation points. Because they are hand-dug, one can be sure that they will not penetrate more than slightly below the lowest expected level to which the groundwater will fall. They will thus truly represent the watertable.

Their location, however, may not always fit into an appropriate network; they may be sited on topographic highs, for example, or their water levels may be deeper than 2 m below the land surface. Another possible disadvantage is that such wells usually have a large diameter. This means that they have a large storage capacity, implying that the water level in the well will take some time to respond to changes in the watertable, or to recover when water is taken from them in substantial quantities. Other causes of erroneous data may be a clogged well screen or a low permeability of the water-transmitting layer.

Relatively deep wells piercing alternating layers of sand and clay below the watertable must be considered with caution; their water levels may be a composite of the different hydraulic heads that occur in the pierced sandy layers.

Before existing wells are included in the network of observation wells, therefore, information should be collected on their depth, diameter, construction, layers penetrated, and frequency of use.

2.4.2 Observation Wells and Piezometers

In addition to properly selected existing wells, a number of watertable observation wells should be placed at strategic points throughout the project area. They may be cased or uncased wells, depending on the stability of the soil at each location.

Uncased Wells

Uncased wells can easily be made with a hand auger as used in soil surveys, and can be 50 to 80 mm in diameter (Figure 2.12A). They can be used successfully in soils

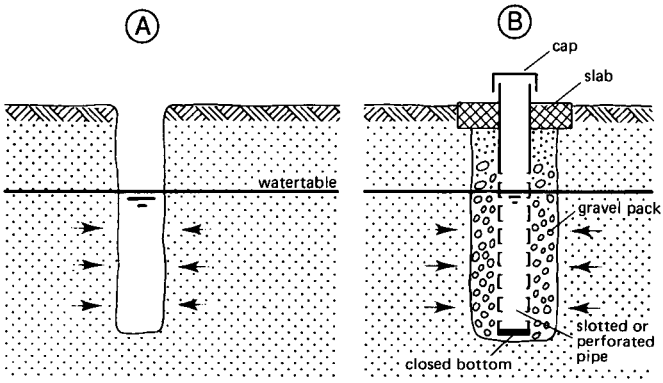

Figure 2.12 Observation wells:
A: Uncased well in stable soil; B: Cased well in unstable soil

that are stable enough to prevent the borehole from collapsing. They are also a cheap means of measuring watertable levels during the first phase of a project (reconnaissance survey), when the primary objective is to obtain a rough idea of the groundwater conditions in the project area.

Cased Wells

When making an observation well in unstable soil, one has to use a temporary casing, say 80 or 100 mm in diameter. The casing prevents sloughing and caving and makes it possible to bore a hole that is deep enough to ensure that it always holds water. Whatever casing material is locally available can be used: sheet metal, drain pipe, or standard commercial types of well casing (steel or PVC).

One starts making a borehole by hand auger or other light-weight boring equipment until one reaches the watertable. After lowering a casing, at least 80 mm in diameter, into the hole, one deepens the hole by bailing out the material inside the casing. When there is difficulty in keeping the sand from heaving inside the casing, this can be overcome by adding water to keep the water level in the pipe above the water level in the water-bearing layer.

When the borehole has reached the required depth, a pipe at least 25 mm in diameter is then lowered inside the casing to the bottom of the hole. Centering this pipe in the casing is important. The pipe's lower end must contain slots or perforations over a length equal to the distance over which the watertable is suspected to fluctuate.

The next step is to fill the space between the pipe and the casing (annular space) with graded coarse sand or fine gravel up to some distance above the upper limit of the slots or perforations; the remaining annular space can be backfilled with parent materials. A properly placed gravel pack facilitates the flow of groundwater into the pipe, and vice versa, and prevents the slots or perforations from becoming clogged by fine particles like clay and silt.

Finally, one pulls out the casing and places a concrete slab around the pipe to protect it from damage (Figure 2.12B).

A gravel pack is not always needed (e.g. if the whole soil profile consists of sand and gravel, free of silt and clay). Wrapping a piece of jute or cotton around the perforated part of the pipe may then suffice. It is advisable to remove any muddy water from the completed well by bailing.

Piezometers

A piezometer is a small-diameter pipe, driven into, or placed in, the subsoil so that there is no leakage around the pipe and all water enters the pipe through its open bottom. Piezometers are particularly useful in project areas where artesian pressures are suspected or in irrigated areas where the rate of downward flow of water has to be determined. A piezometer indicates only the hydrostatic pressure of the groundwater at the specific point in the subsoil at its open lower end.

In a partly saturated homogeneous sand layer, vertical flow components are usually lacking or are of such minor importance that they can be neglected. Hence, at any depth in such a layer, the hydraulic head corresponds to the watertable height; in other words, in measuring the watertable, it makes no difference how far the piezometer penetrates into the sand layer, as is shown in Figure 2.13A. In such cases, a single piezometer will suffice.

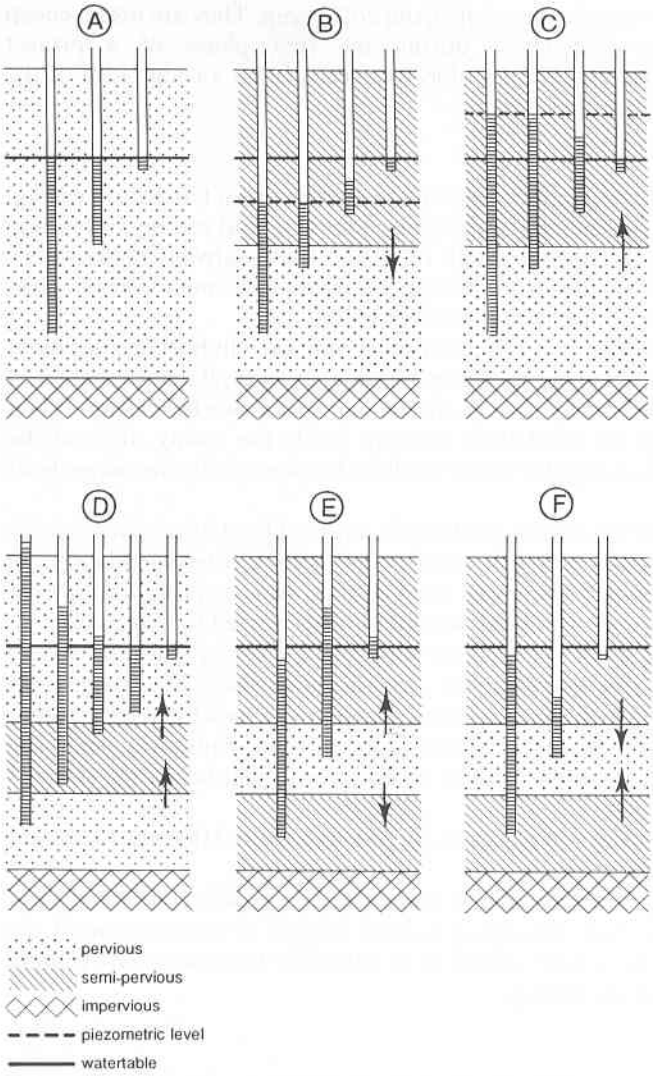

Figure 2.13 Examples of water levels in piezometers for different conditions of soil and groundwater

The same applies to a fully saturated confined sand layer. As it is generally assumed that the flow of groundwater through such a layer is essentially horizontal and that vertical flow components can be neglected, the distribution of hydraulic head in the layer is the same everywhere in the vertical plane. It suffices therefore to place only one piezometer in such a layer. Its water level is known as the hydraulic head of the layer, or the piezometric head or the potential head (Chapter 7).

In stratified soils, piezometers are useful in determining whether the groundwater is moving upward or downward. They are also useful in determining whether any natural drainage occurs in the project area. The piezometers of Figure 2.13B and F

indicate that there is natural drainage through the sand layer.

Since the flow of groundwater through confining layers (clay, loamy clay, clay loam, silty clay loam) is mainly vertical, the water level in a piezometer that penetrates into such a layer is a function of its depth of penetration (Figure 2.13B, C, and D).

Piezometers can be installed by driving or jetting them into position with a high-velocity water jet. If more than a few piezometers are to be installed, the jetting technique is recommended. Although this technique is fast, a disadvantage is that it does not provide precise information on the pierced materials, unless the piezometers are installed at the location of a borehole whose log is available. Another method of installing piezometers is to make a borehole 100 to 200 mm in diameter and then install three or four piezometers at different depths (Figure 2.14). To prevent leakage, care should be taken that pierced clay layers are properly sealed.

The lower end of a piezometer can easily become clogged by fine materials that enter the pipe. This can be avoided by perforating the lower 0.3 to 0.5 m of the pipe. To prevent fine soil particles from clogging the tiny holes, some jute or cotton can be wrapped around the perforations and the lower open end sealed with a plug. Graded coarse sand or fine gravel placed around the perforated part of the pipe will facilitate the flow of water into the pipe and vice versa. Strictly speaking, a perforated pipe cannot be called a piezometer, but because the perforations cover only the lower few decimetres of the pipe, we shall retain the term piezometer.

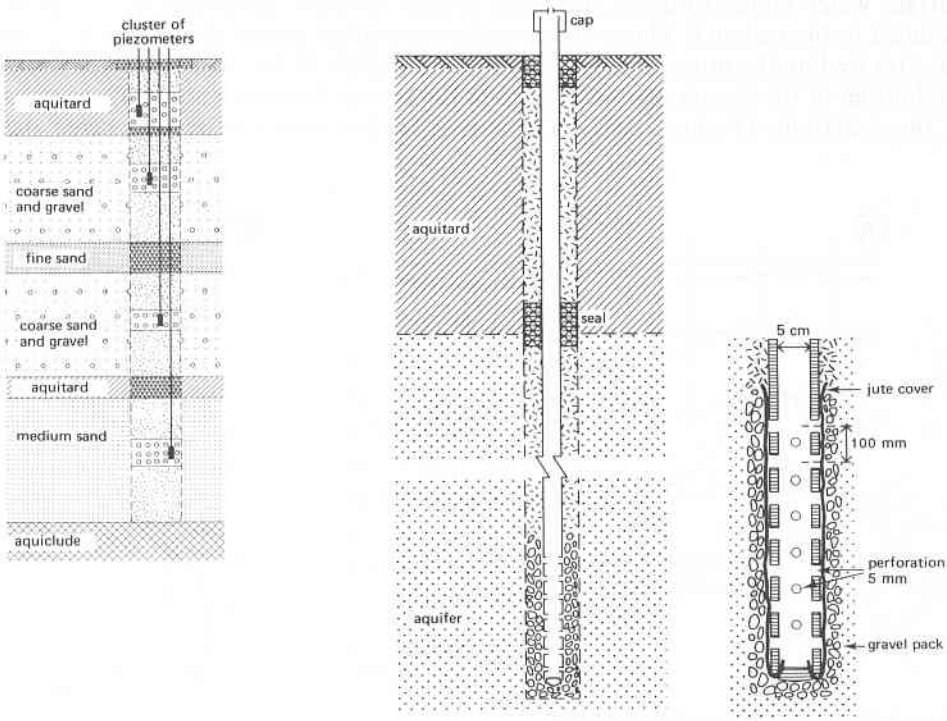

Figure 2.14 Multiple piezometer well and the cross-section of a piezometer

2.4.3 Observation Network

Layout

To save costs, observation wells and piezometers should be installed concurrently with soil borings that are needed to explore the shallow subsurface. These borings are usually made on a rectangular grid pattern that is laid out on the basis of information on topography, geology, soils, and hydrology collected during the early phase of the project. Figure 2.15 shows some examples of grid systems.

Soil borings should be spaced rather close together to make it possible to correlate subsurface layers. It is not necessary to transform each soil boring into an observation well because the watertable is a smooth surface. Nevertheless, abrupt changes in the configuration of the watertable do occur, due to discontinuities of soil layers, outflow of groundwater into streams, pumping from wells, and local irrigation. So, in planning a network of observation wells, one should note that they will be required:

- Along, and perpendicular to, lines of suspected groundwater flow;
- At locations where changes in the slope of the watertable occur or are suspected;
- On the banks of streams or other open water courses and along lines perpendicular to them;
- In areas where shallow watertables occur or can be expected in the future (areas with artesian pressure and areas with a high intensity of irrigation);
- Along and perpendicular to the (project) area's boundaries.

Surface water bodies in direct hydraulic contact with the groundwater should be included in the network. These surface waters are either fed by the groundwater or they are feeding the groundwater (Figure 2.16A and B). If the watertable lies below the bottom of the stream, the water level of the stream does not represent any point of the watertable (Figure 2.16C). The stream is then losing water that percolates

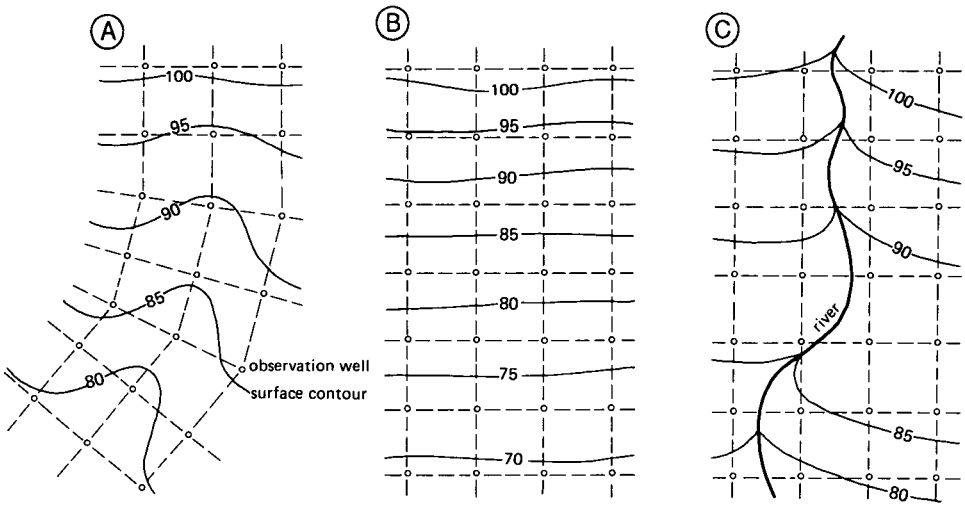

Figure 2.15 Different layouts of grid systems:
A: For a narrow valley; B: For a uniformly sloping area; C: For an almost level alluvial plain

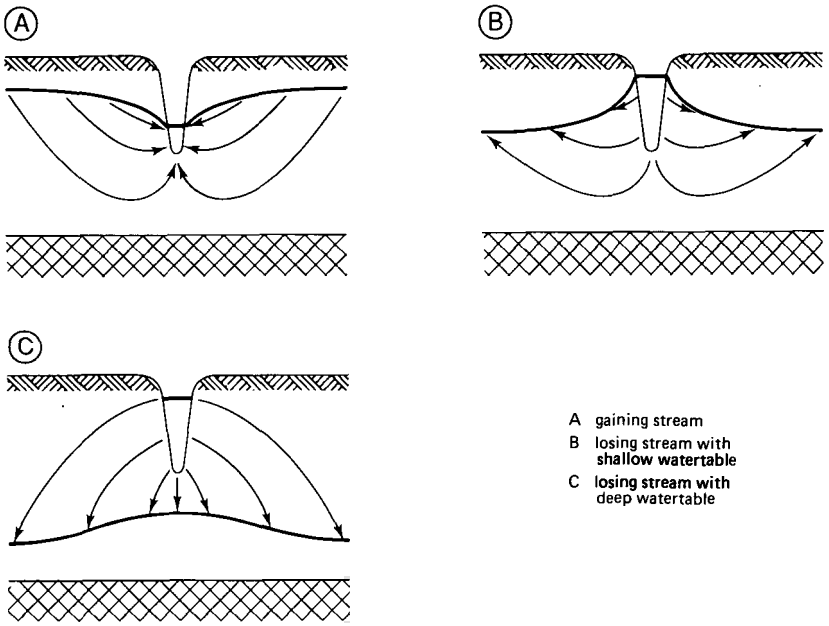

Figure 2.16 Gaining and losing streams

through the unsaturated zone to the deep watertable. A local mound is built up under the stream and its height can be measured by placing an observation well on the bank of the stream.

Water levels of streams or other water courses represent local mounds or depressions in the watertable and consequently are of great importance in a study of groundwater conditions. At strategic places in these water courses, therefore, staff gauges should be installed.

Density

No strict rule can be given as to the density of the observation network, because this depends entirely on the topographic, geological, and hydrological conditions of the area, and on the type of survey (reconnaissance, semi-detailed, detailed). As the required accuracy is generally inversely proportional to the size of the area, the relation given in Table 2.1 may serve as a rough guide.

In areas where the subsurface geology is fairly uniform, the watertable is usually smooth and there will be no abrupt changes. In such areas, the observation wells can be spaced farther apart than in areas where the subsurface geology is heterogeneous. Near lines of recharge or discharge (e.g. streams or canals), the spacing of the wells could be decreased in approximately the following sequence: 1000, 500, 250, 100, 40, 15, 5 m.

Depth

The depth of observation wells should be based on the expected lowest groundwater level. This will ensure that the wells do not fall dry in the dry season and that readings

Table 2.1 Relation between size of area and number of observation points

Size of area under study (ha)	No. of observation points	No. of observation points per 100 ha
100	20	20
1000	40	4
10 000	100	1
100 000	300	0.3

can be taken throughout a full hydrological year. The lowest water level can only be estimated, unless data from previous investigations are available. Generally, watertables deeper than about 3 m are not interesting from the viewpoint of planning a drainage system. Observation wells to this depth are therefore adequate in most flat lands. In areas with a rolling topography, deeper observation wells may be needed on the topographic highs to obtain a complete picture of the groundwater conditions.

In stratified soils and particularly in areas where artesian pressure exists or can be suspected, a number of deep piezometers are needed in addition to the shallow ones. No rule can be given as to how many of these should be placed or how deep they should be, because this depends on the hydrogeological conditions in the area. It is a matter of judgement as the investigations proceed. In profiles as shown in Figure 2.13B and C, double piezometer wells may suffice: one in the covering low-permeable layer and the other in the underlying sand layer.

In many alluvial plains, the covering layer is made of alternating layers of heavy and light-textured materials, or even peat. The total thickness of this layer can be many metres. In this case, a multipiezometer well can best be made, containing 3 to 5 piezometers placed at different depths (e.g. at 2, 5, 8, 12, and 15 m). The deepest piezometer should be placed in the underlying coarse sand layer, where groundwater is under artesian pressure. It will be obvious that, in making such wells, one will need power augers or jetting equipment.

Well Elevations

To determine the elevation of the observation wells and piezometers and thus be able to correlate watertable levels with land surface levels, a levelling survey must be made.

Water levels in the wells are measured from a fixed measuring point, which, for cased wells, can be the rim of the casing; for uncased wells, a measuring point must be made (e.g. a piece of steel, wood, or stone).

2.4.4 Measuring Water Levels

Methods

Water-level measurements can be taken in various ways (Figure 2.17):

- The wetted tape method (Figure 2.17A): A steel tape (calibrated in millimetres), with a weight attached to it, is lowered into the pipe or borehole to below the water

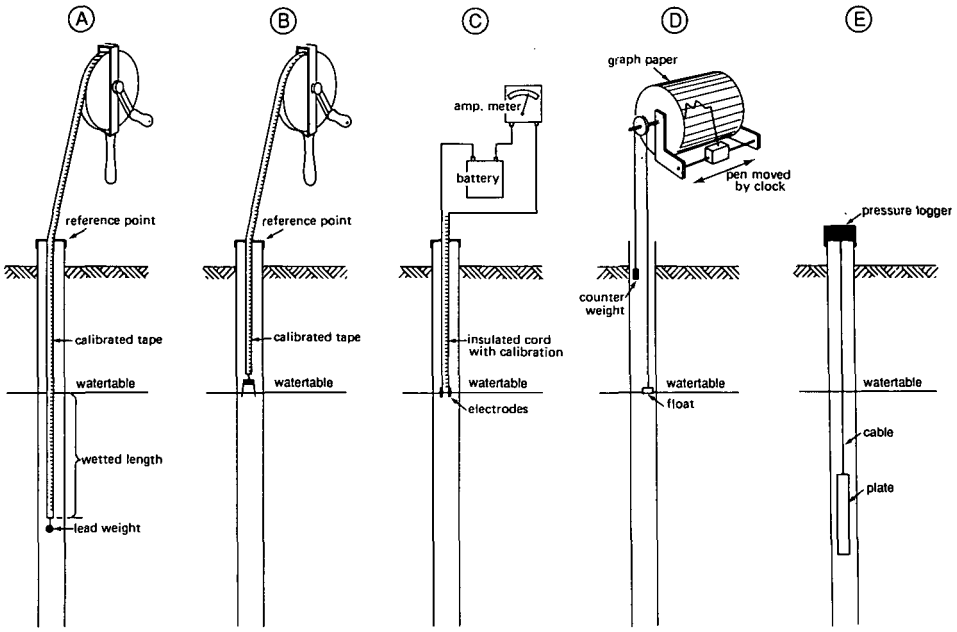

Figure 2.17 Various ways of measuring depth to water level in wells or piezometers

level. The lowered length of tape from the reference point is noted. The tape is then pulled up and the length of its wetted part is measured. (This is facilitated if the lower part of the tape is chalked.) When the wetted length is subtracted from the total lowered length, this gives the depth to the water level below the reference point;

- With a mechanical sounder (Figure 2.17B): This consists of a small steel or copper tube (10 to 20 mm in diameter and 50 to 70 mm long), which is closed at its upper end and connected to a calibrated steel tape. When lowered into the pipe, it produces a characteristic plopping sound upon hitting the water. The depth to the water level can be read directly from the steel tape;
- With an electric water-level indicator (Figure 2.17C): This consists of a double electric wire with electrodes at their lower ends. The upper ends of the wire are connected to a battery and an indicator device (lamp, mA meter, sounder). When the wire is lowered into the pipe and the electrodes touch the water, the electrical circuit closes, which is shown by the indicator. If the wire is attached to a calibrated steel tape, the depth to the water level can be read directly;
- With a floating level indicator or recorder (Figure 2.17D): This consists of a float (60 to 150 mm in diameter) and a counterweight attached to an indicator or recorder. Recorders can generally be set for different lengths of observation period. They require relatively large pipes. The water levels are either drawn on a rotating drum or punched in a paper tape;
- With a pressure logger or electronic water-level logger (Figure 2.17E): This measures and records the water pressure at one-hour intervals over a year. The pressure recordings are controlled by a microcomputer and stored in an internal, removable memory block. At the end of the observation period or when the memory block

has reached capacity, it is removed and replaced. The recorded data are read by a personal computer. Depending on the additional software chosen, the results can be presented raw or in a calculated form. Pressure loggers have a small diameter (20 to 30 mm) and are thus well suited for measurements in small-diameter pipes;
- The water levels of open water surfaces are usually read from a staff gauge or a water-level indicator installed at the edge of the water surface. A pressure logger is most convenient for this purpose, because no special structures are required; the cylinder need only be anchored in the river bed.

Frequency of Measurements
The watertable reacts to the various recharge and discharge components that characterize a groundwater system and is therefore constantly changing. Important in any drainage investigation are the (mean) highest and the (mean) lowest watertable positions, as well as the mean watertable of a hydrological year. For this reason, water-level measurements should be made at frequent intervals for at least a year. The interval between readings should not exceed one month, but a fortnight may be better. All measurements should, as far as possible, be made on the same day because this gives a complete picture of the watertable.

Each time a water-level measurement is made, the data should be recorded in a notebook. It is advisable to use pre-printed forms for this purpose. An example is shown in Figure 2.18. Even better is to enter the data in a computerized database system. Recorded for each observation are: date of observation, observed depth of the water level below the reference point, calculated depth below ground surface (for free watertables only), and calculated water-level elevation (with respect to a general datum plane, e.g. mean sea level). Other particulars should also be noted (e.g. number of the well, its location, depth, surface elevation, reference point elevation).

If one wants to study the effect that a rainshower or an irrigation application has on the watertable, daily or even continuous readings may be needed. A pressure logger or an automatic recorder should then be installed in a representative large-diameter well; depending upon the type of recorder selected the well should have a certain minimum diameter, e.g. 7 cm.

2.4.5 Groundwater Quality

For various reasons, a knowledge of the groundwater quality is required. These are:
- Any lowering of the watertable may provoke the intrusion of salty groundwater from adjacent areas, or from the deep underground, or from the sea. The drained area and its surface water system will then be charged daily with considerable amounts of dissolved salts;
- The disposal of the salty drainage water into fresh-water streams may create environmental and other problems, especially if the water is used for irrigation and/ or drinking;
- In arid and semi-arid regions, soil salinization is directly related to the depth of the groundwater and to its salinity;
- Groundwater quality dictates the type of cement to be used for hydraulic structures, especially when the groundwater is rich in sulphates;

GENERAL DATA OBSERVATION POINT NO.....

MAP NO. ..... COORDINATES: X = ... Y = ...

MUNICIPALITY ..... PROVINCE .....

OWNER ..... INSTALLATION DATE ..... TYPE[1] .....

DEPTH ..... SCREENED PART ..... AQUIFER TYPE[2] .....

WELL LOG: FILE NO. ..... WATER SAMPLES: FILE NO. .....

SURFACE ELEVATION ..... REFERENCE POINT ELEVATION .....

OBSERVATIONS

DATE	READING[3]	ELEVATION[4]	DEPTH[5]	REMARKS[6]

1 *e.g. village well, open borehole, piezometer*
2 *e.g. unconfined aquifer, semi-confined aquifer, semi-pervious covering layer*
3 *with respect to reference point*
4 *with respect to general datum, for example mean sea level*
5 *below ground surface (for phreatic levels only)*
6 *data on water sample, irrigation, water at the surface, flow from wells, water withdrawal (pumping), etc.*

Figure 2.18 Example of a form for recording water levels

- Agricultural crops are affected by groundwater quality if the groundwater approaches the rootzone.

Sources of Salinity

All groundwater contains salts in solution. The type of salts depends on the geological environment, the source of the groundwater, and its movement. The weathering of primary minerals is the direct source of salts in groundwater. Bicarbonate (HCO_3)

is usually the primary anion in groundwater and forms as a result of the solution of carbon dioxide in water. Carbon dioxide is a particularly active weathering agent for such source rocks as limestone and dolomite.

Sodium in the water originates from the weathering of feldspars (albite), clay minerals, and the solution of evaporites (halite and mirabilite). Evaporites are also the major natural source of chloride in groundwater, while sulphate originates from the oxidation of sulphide ores or the solution of gypsum and anhydrite. Such primary minerals as amphiboles (hornblende), apatite, fluorite, and mica are the sources of fluoride in groundwater. The mineral tourmaline is the source of boron.

In the groundwater of coastal and delta plains, the sea is the source of salinity.

Groundwater quality is also related to the relief of the area. Fresh groundwater usually occurs in topographic highs which, if composed of permeable materials, are areas of recharge. On its way to topographic lows (areas of discharge), the groundwater becomes mineralized through the solution of minerals and ion exchange. Groundwater salinity varies with the texture of sediments, the solubility of minerals, and the contact time. Groundwater salinity tends to be highest where the movement of the groundwater is least, so salinity usually increases with depth.

Irrigation also acts as a source of salts in groundwater. It not only adds salts to the soil, but also dissolves salts in the root zone. Water that has passed through the root zone of irrigated land usually contains salt concentrations several times higher than that of the originally applied irrigation water.

Evapotranspiration tends to concentrate the salinity of groundwater. Highly saline groundwater can therefore be found in arid regions with poor natural drainage and consequently a shallow watertable.

The choice of a method for measuring groundwater salinity depends on the reason for making the measurements, the size of the area – and thus the number of samples to be taken and measured – and the time and the budget available for doing the work.

Once the network of observation wells, boreholes, and piezometers has been established, water samples should be taken in a representative number of them. Sampling can often best be combined with other drainage investigations, such as measuring hydraulic conductivity in open boreholes. The sample is then taken after a sufficient quantity of water has been bailed from the hole.

Electrical Conductivity

A rapid determination of the salinity of groundwater can be made by measuring its electrical conductivity, EC. Conductivity is preferred rather than its reciprocal, resistance, because the EC increases with the salt content. Electrical conductivity defines the conductance of a cubic centimetre of water at a standard temperature of 25°C. It is expressed in decisiemens per metre (dS/m), formerly in millimhos per centimetre (mmhos/cm). Expressing the results in terms of specific electrical conductivity makes the determination independent of the size of the water sample. Conductivity cannot simply be related to the total dissolved solids because groundwater contains a variety of ionic and undissociated species. An approximate relation for most groundwater with an EC-value in the range of 0.1 to 5 dS/m is: 1 dS/m $\approx$ 640 mg/l (Chapter 15).

Major Chemical Constituents
The EC expresses the total concentration of soluble salts in the groundwater, but gives no information on the types of salts. Needed for this purpose are laboratory determinations of such constituents as calcium, magnesium, sodium, carbonate, bicarbonate, chloride, sulphate, and nitrate. Since these chemical analyses are costly, not all the observation points need be sampled for detailed analysis. A selection of sites should be made, based on the results of the EC measurements.

For more information on groundwater quality reference is made to Hem 1970.

2.5 Processing the Groundwater Data

Before any conclusions can be drawn about the cause, extent, and severity of an area's drainage problems, the raw groundwater data on water levels and water quality have to be processed. They then have to be related to the geology and hydrogeology of the area. The results, presented in graphs, maps, and cross-sections, will enable a diagnosis of the problems to be made.

We shall assume that such basic maps as topographic, geological, and pedological maps are available.

The following graphs and maps have to be prepared that are discussed hereunder:
- Groundwater hydrographs;
- Watertable-contour map;
- Depth-to-watertable map;
- Watertable-fluctuation map;
- Head-differences map;
- Groundwater-quality map.

2.5.1 Groundwater Hydrographs

When the amount of groundwater in storage increases, the watertable rises; when it decreases, the watertable falls. This response of the watertable to changes in storage can be plotted in a hydrograph (Figure 2.19). Groundwater hydrographs show the water-level readings, converted to water levels below ground surface, against their corresponding time. A hydrograph should be plotted for each observation well or piezometer.

In land drainage, it is important to know the rate of rise of the watertable, and even more important, that of its fall. If the groundwater is not being recharged, the fall of the watertable will depend on:
- The transmissivity of the water-transmitting layer, KH;
- The storativity of this layer, S;
- The hydraulic gradient, dh/dx.

After a period of rain (or irrigation) and an initial rise in groundwater levels, they then decline, rapidly at first, and then more slowly as time passes because both the hydraulic gradient and the transmissivity decrease. The graphical representation of

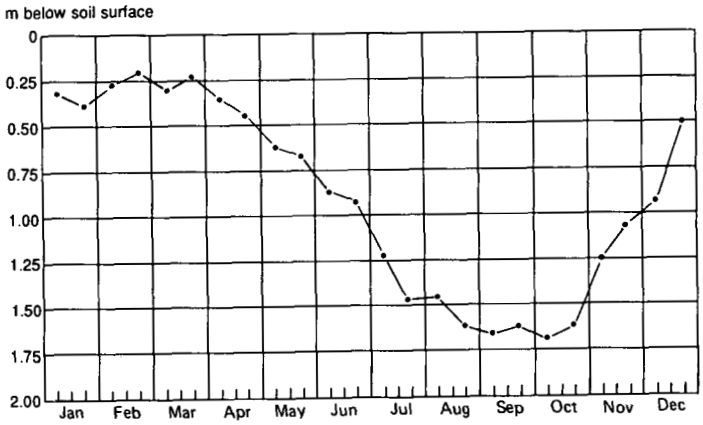

Figure 2.19 Hydrograph of a watertable observation well

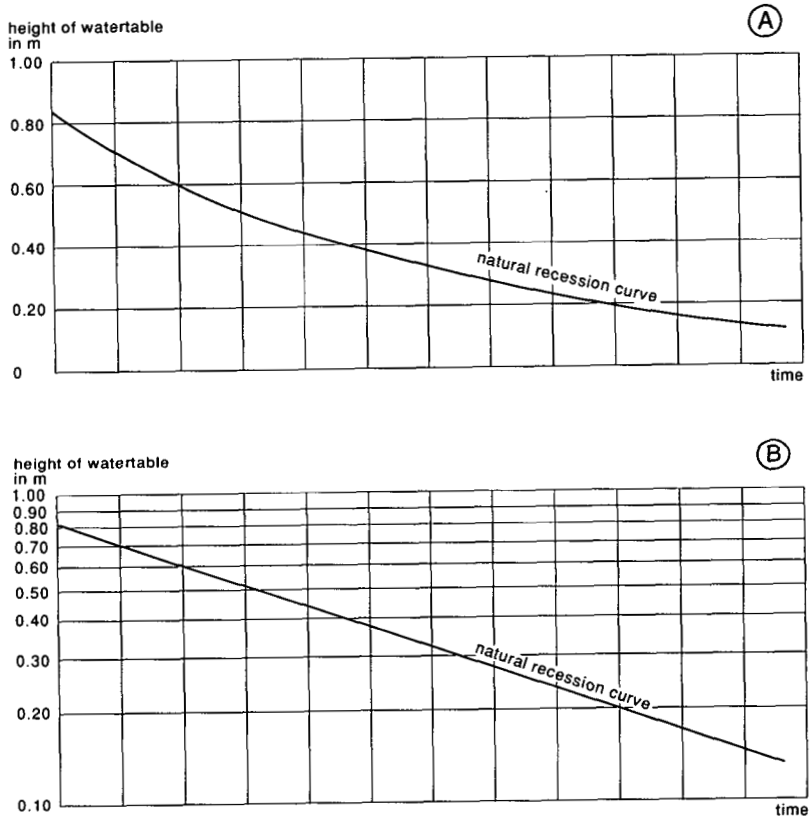

Figure 2.20 Natural recession of groundwater level:
A: Linear scale; B: Logarithmic scale

the watertable decline is known as the natural recession curve. It can be shown that the logarithm of the watertable height decreases linearly with time. Hence, a plot of the watertable height against time on semi-logarithmic paper gives a straight line (Figure 2.20). Groundwater recession curves are useful in studying changes in groundwater storage and in predicting future groundwater levels.

2.5.2 Groundwater Maps

Watertable-Contour Map

A watertable-contour map shows the elevation and configuration of the watertable on a certain date. To construct it, we first have to convert the water-level data from the form of depth below surface to the form of watertable elevation (= water level height above a datum plane, e.g. mean sea level). These data are then plotted on a topographic base map and lines of equal watertable elevation are drawn. A proper contour interval should be chosen, depending on the slope of the watertable. For a flat watertable, 0.25 to 0.50 m may suit; in steep watertable areas, intervals of 1 to 5 m or even more may be needed to avoid overcrowding the map with contour lines.

The topographic base map should contain contour lines of the land surface and should show all natural drainage channels and open water bodies. For the given date, the water levels of these surface waters should also be plotted on the map. Only with these data and data on the land surface elevation can watertable contour lines be drawn correctly (Figure 2.21).

To draw the watertable-contour lines, we have to interpolate the water levels between the observation points, using the linear interpolation method as shown in Figure 2.22.

Instead of preparing the map for a certain date, we could also select a period (e.g. a season or a whole year) and calculate the mean watertable elevation of each well for that period. This has the advantage of smoothing out local or occasional anomalies in water levels.

A watertable-contour map is an important tool in groundwater investigations because, from it, one can derive the gradient of the watertable (dh/dx) and the direction

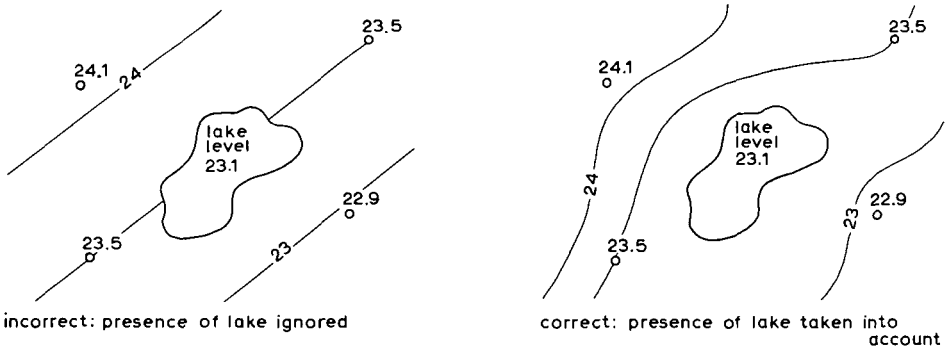

Figure 2.21 Example of watertable-contour lines
A: Incorrectly drawn; B: Correctly drawn

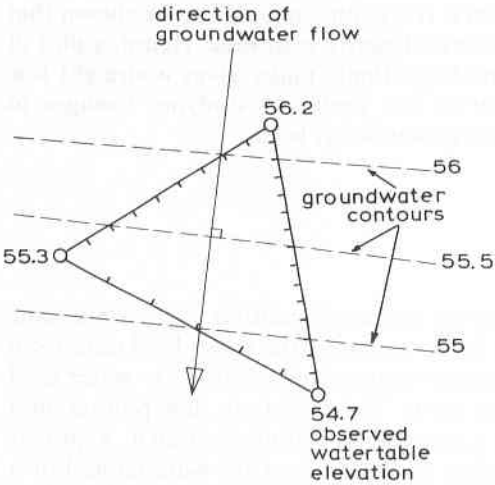

Figure 2.22 Construction of watertable-contour lines by linear interpolation

of groundwater flow, which is perpendicular to the watertable-contour lines.

Figure 2.23A presents an example of a topographical base map of an irrigated area with its grid system of observation points; Figure 2.23B shows the watertable contour map.

For artesian or irrigation areas in which two or more piezometers have been installed at the same location, with the bottom of each at a different depth in a different soil layer (as in Figure 2.14), contour maps of the hydraulic head in each layer should be made.

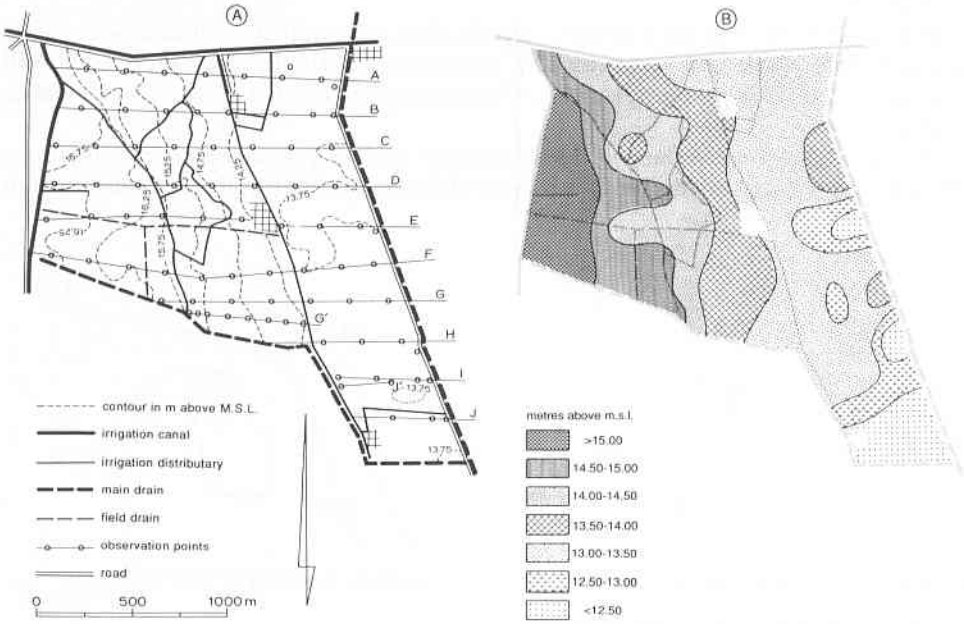

Figure 2.23 A: Topographic base map of an irrigated area; B: Its watertable contour map

Depth-to-Watertable Map

A depth-to-watertable map, or isobath map, as these names imply, shows the spatial distribution of the depth of the watertable below the land surface. It can be prepared in two ways. The water-level data from all observation wells for a certain date are first converted to water levels below land surface. (The reference point from which the readings were taken need not necessarily be the land surface.) One then plots the converted data on a topographic base map and draws isobaths or lines of equal depth to groundwater (Figure 2.24). A suitable contour interval could be 0.50 m.

The other way of preparing this map is by superimposing a watertable-contour map made for a certain date on the topographic base map showing contour lines of the land surface. From the two families of contour lines, the difference in elevation at contour intersections can be read. These data are then plotted on a clean topographic map, and the isobaths are drawn.

Depth-to-watertable maps are usually prepared for critical dates (e.g. when farming operations have to be performed or when the crops are expected to be most sensitive to high watertables). Instead of preparing the isobath map for a special date, one can also choose a season and prepare a map showing the mean depth-to-watertable for that season. Periods or seasons during which the watertables are highest and lowest can be read from groundwater hydrographs (Figure 2.19)

Watertable-Fluctuation Map

A watertable-fluctuation map is a map that shows the magnitude and spatial distribution of the change in watertable over a period (e.g. a whole hydrological year). Using such graphs as shown in Figure 2.19, we calculate the difference between the highest and the lowest watertable height (or preferably the difference between the mean highest and the mean lowest watertable height for the two seasons). We then plot these data on a topographic base map and draw lines of equal change in watertable, using a convenient contour interval.

A watertable-fluctuation map is a useful tool in the interpretation of drainage

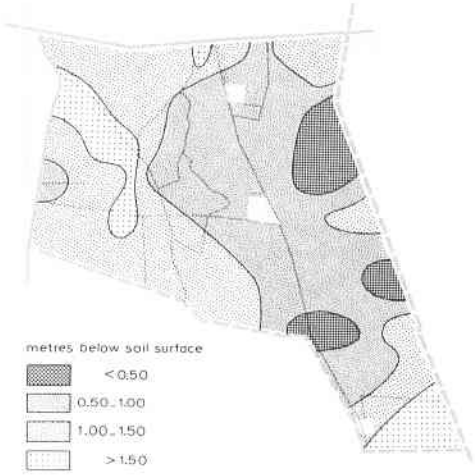

Figure 2.24 Example of a depth-to-watertable (or isobath) map

problems in areas with large watertable fluctuations, or in areas with poor natural drainage (or upward seepage) and permanently high watertables (i.e. areas with minor watertable fluctuations).

Head-Differences Map

A head-differences map is a map that shows the magnitude and spatial distribution of the differences in hydraulic head between two different soil layers. Let us assume a common situation as shown in Figure 2.13B or 2.13C. We then calculate the difference in water level between the shortest piezometer and the longest, and plot the result on a map. After choosing a proper contour interval (e.g. 0.10 or 0.20 m), we draw lines of equal head difference.

Another way of drawing such a map is to superimpose a watertable-contour map on a contour map of the piezometric surface of the underlying layer. We then read the head differences at contour line intersections, plot these on a base map, and draw lines of equal head difference. The map is a useful tool in estimating upward or downward seepage.

Groundwater-Quality Map

A groundwater-quality or electrical-conductivity map is a map that shows the magnitude and spatial variation in the salinity of the groundwater. The EC values of all representative wells (or piezometers) are used for this purpose (Figure 2.25).

Groundwater salinity varies not only horizontally but also vertically; a zonation of groundwater salinity is common in many areas (e.g. in delta and coastal plains, and in arid plains). It is therefore advisable to prepare an electrical-conductivity map not only for the shallow groundwater but also for the deep groundwater.

Other types of groundwater-quality maps can be prepared by plotting different quality parameters (e.g. Sodium Adsorption Ratio (SAR) values; see Chapter 15).

Figure 2.25 Example of a groundwater-quality map of shallow groundwater

2.6 Interpretation of Groundwater Data

It must be emphasized that a proper interpretation of groundwater data, hydrographs, and maps requires a coordinate study of a region's geology, soils, topography, climate, hydrology, land use, and vegetation. If the groundwater conditions in irrigated areas are to be properly understood and interpreted, cropping patterns, water distribution and supply, and irrigation efficiencies should be known too.

2.6.1 Interpretation of Groundwater Hydrographs

Watertable changes are of two kinds:
- Changes due to changes in groundwater storage;
- Changes due to other influences (e.g. changes in atmospheric pressure, deformation of the water-transmitting layer, earthquakes).

In drainage studies, we are primarily interested in watertable changes due to changes in groundwater storage because they are the result of the groundwater regime (i.e. the way by which the groundwater is recharged and discharged). Rising watertables indicate the periods when recharge is exceeding discharge and falling watertables the periods when discharge is exceeding recharge (Figure 2.26).

Rather abrupt changes in the amount of water stored in the subsoil will be found in land adjacent to stream channels because that land will be influenced by the rise and fall of the stream stage (Figure 2.27), and in areas of relatively shallow watertables influenced by precipitation or irrigation.

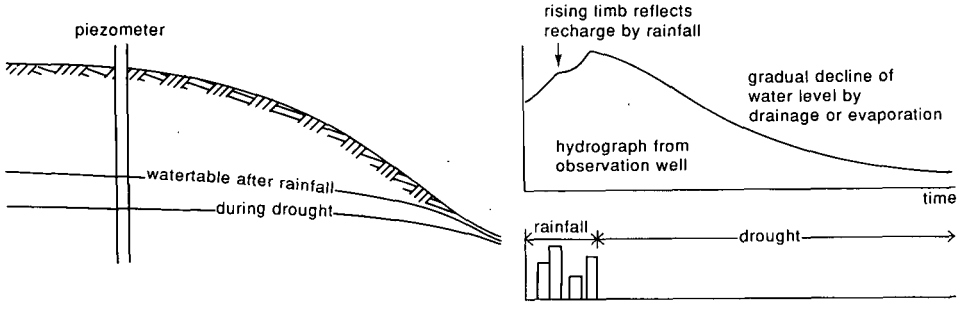

Figure 2.26 Groundwater hydrograph showing a rise of the watertable during recharge by rain and its subsequent decline during drought

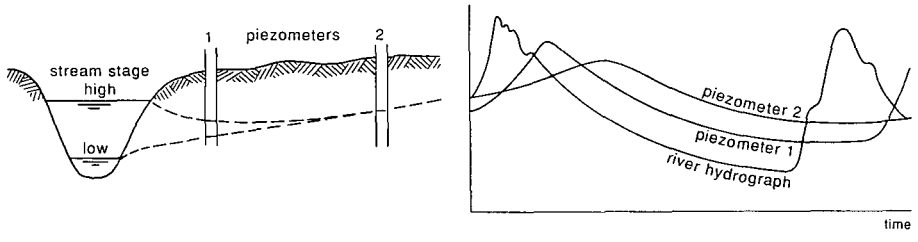

Figure 2.27 Influence of the stream stage on the watertable in adjacent land. Note that the influence diminishes with increasing distance from the stream

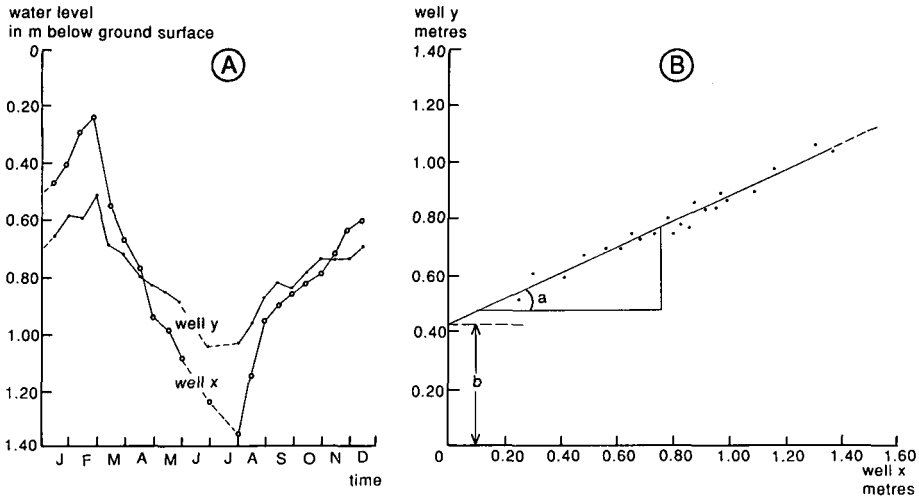

Figure 2.28 A: Hydrographs of observation wells x and y; B: Correlation of the water levels in these wells, showing the regression line obtained by a linear regression of y upon x

Although the effect of precipitation on the watertable is usually quite clear, an exact correlation often poses a problem because:

- Differences in drainable pore space of the soil layers in which the watertable fluctuates will cause the watertable to rise or fall unevenly;
- Part of the precipitation may not reach the watertable at all because it evaporates or because it is discharged as surface runoff and/or is stored in the zone of aeration (soil water);
- The groundwater-flow terms may result in a net recharge or a net discharge of the groundwater, thus affecting the watertable position.

The groundwater hydrographs of all the observation points should be systematically analyzed. A comparison of these hydrographs enables us to distinguish different groups of observation wells. Each well belonging to a certain group shows a similar response to the recharge and discharge pattern of the area. By a similar response, we mean that the water level in these wells starts rising at the same time, attains its maximum value at the same time, and, after recession starts, reaches its minimum value at the same time. The amplitude of the water level fluctuation in the various wells need not necessarily be exactly the same, but should show a great similarity (Figure 2.28A). Areas where such wells are sited can then be regarded as hydrological units (i.e. sub-areas in which the watertable reacts to recharge and discharge everywhere in the same way).

The water-level readings of a certain well in a group of wells can be correlated with those of another well in that group, as is shown in Figure 2.28B. To calculate the correlation of two wells, the method of linear regression is used (Chapter 6). If the two wells correlate satisfactorily, one of the two can be dropped from the network. Such an analysis may lead to the selection of a number of standard observation wells only, and the network can thus be reduced. From the water-level readings in these standard wells, which form the base network, the water levels in the other observation

Table 2.2 Hydrological sub-areas with their mean depths to groundwater for the wet and dry seasons, in m below soil surface

	Hydrological sub-area (groundwater depth group)					
	A	B	C	D	E	F
Mean depth to groundwater in the wet season	0.30	0.45	0.60	0.80	1.10	1.90
Mean depth to groundwater in the dry season	0.60	0.80	1.00	1.20	1.50	2.40

wells that were dropped can be calculated from the established regression equation.

For further evaluation of the groundwater conditions in each hydrological sub-area, we can calculate the mean depth to groundwater for the wet season and that for the dry season, using the water-level measurements of all wells in the sub-areas. Table 2.2 shows an example of such a grouping of levels.

Figure 2.29 shows the depth to groundwater for the hydrological sub-areas A, D, and F in an experimental field of sandy soils in the eastern part of The Netherlands over the period 1961 to 1967 (Colenbrander 1970).

Sub-area A is a typical seepage area characterized by shallow watertables that are influenced by a seasonal precipitation surplus. Sub-area F is a typical area with good natural drainage; seasonal rains do cause the watertable to rise, but seldom higher than 1.50 m below the ground surface. Sub-area D takes a somewhat intermediate position between the other two; the mean depths to groundwater in the wet and dry

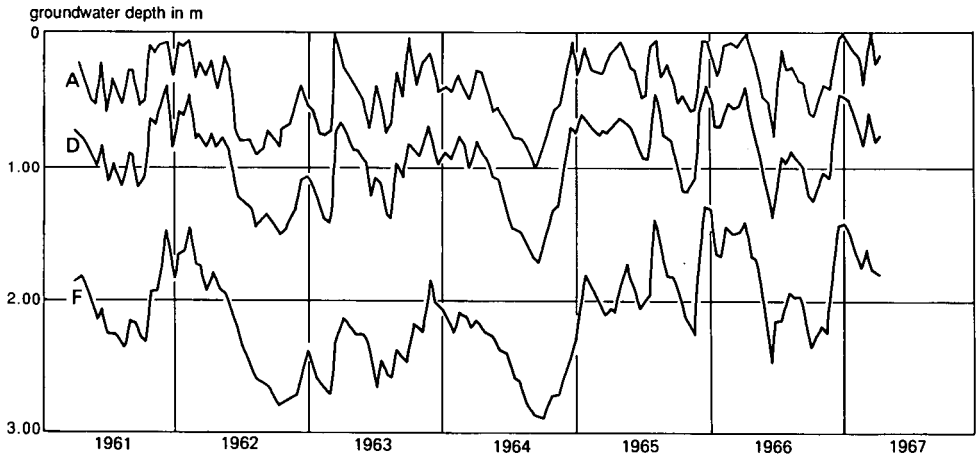

Figure 2.29 Mean depth to watertable in three homogeneous hydrological sub-areas over the period 1961 to 1967 (after Colenbrander 1970)

seasons (0.80 and 1.20 m, respectively) do not pose special problems to agriculture, but an incidental precipitation surplus during the wet season may cause the watertable to rise to about 0.50 m or less below the ground surface.

Variations in stream flow are closely related to the groundwater levels in land adjacent to a stream. Stream flow originating from groundwater discharge is known as groundwater runoff or base flow. During fair-weather periods, all stream flow may be contributed by. base flow. To estimate the base flow from an area with fairly homogeneous hydrogeological conditions, the mean groundwater levels of the area are plotted against the stream flow during periods when all flow originates from groundwater. We thus obtain a rating curve of groundwater runoff for the area in question (Figure 2.30).

Groundwater hydrographs also offer a means of estimating the annual groundwater recharge from rainfall. This, however, requires several years of records on rainfall and watertables. An average relationship between the two can be established by plotting the annual rise in watertable against the annual rainfall (Figure 2.31). Extending the straight line until it intersects the abscis gives the amount of rainfall below which there is no recharge of the groundwater. Any quantity less then this amount is lost by surface runoff and evapotranspiration.

Percolating rainwater is not the only reason why watertables fluctuate. Daily fluctuations of the watertable may also be observed in coastal areas due to the tides. Sinusoidal fluctuations of groundwater levels in such areas occur in response to tides.

Finally, changes in atmospheric pressure produce fluctuations in water levels of wells that penetrate confined waterbearing layers. An increase in atmospheric pressure produces a decline of the water level, and vice versa. This phenomenon is due to the

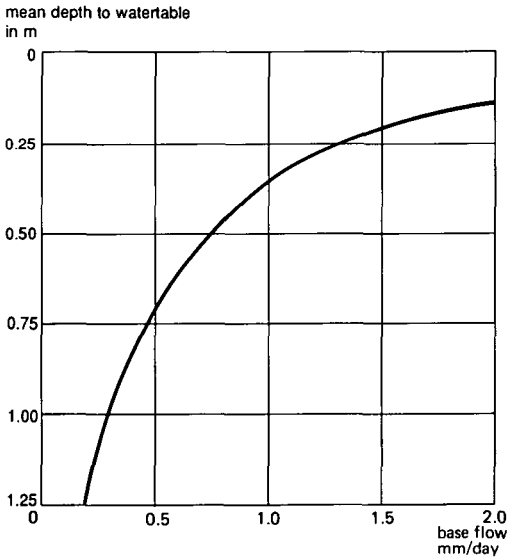

Figure 2.30 Rating curve: Relationship between mean depth to watertable and groundwater runoff (base flow)

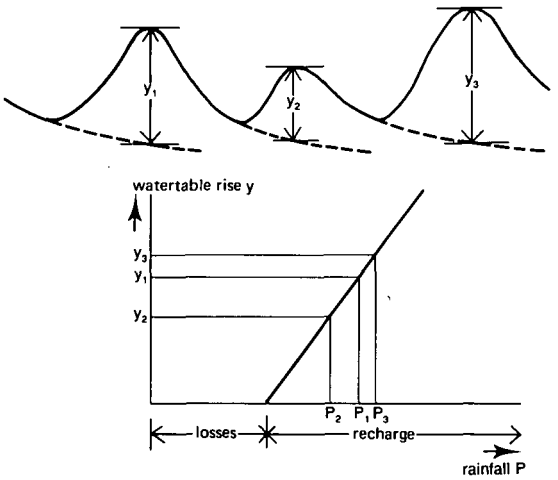

Figure 2.31 Relationship between annual groundwater recharge and rainfall

elasticity of these waterbearing layers.

Continuous or intermittent pumping from wells produces changes in the watertable (or piezometric surface) in the vicinity of such wells. This will be further elaborated in Chapter 10.

2.6.2 Interpretation of Groundwater Maps

Watertable-Contour Maps

Contour maps of the watertable are graphic representations of the relief and slope of the watertable. They are the basis for determining the direction and rate of groundwater flow, the drainage of groundwater from all sources, and the variations in percolation rates and in the permeability of the alluvial materials.

Under natural conditions, the watertable in homogeneous flat areas has little relief and is generally sloping smoothly and gently to low-lying zones of groundwater discharge. In most areas, however, minor relief features in the watertable are common; they consist of local mounds or depressions that may be natural or man-made (Figure 2.32). Groundwater flow is always in the downslope direction of the watertable and, if permeability is assumed to be constant, the fastest movement and largest quantity of groundwater flow are in the direction of maximum slope.

A local mound in the watertable may be due to local recharge of the groundwater by irrigation or by upward seepage. Local depressions may be due to pumping from wells or to downward seepage. Upward or downward seepage is common in alluvial plains underlain by karstic limestones. Buried sinkholes and karstic channels in the limestone are usually sites of recharge or discharge of the overlying alluvial deposits.

The topography of the area under study is important because it controls the configuration of the watertable. The shape of the watertable can be convex or concave. In an area dissected by streams and natural drainage channels, the watertable between adjacent streams (i.e. the interfluves) is convex. In the vicinity of a losing stream, it is concave (Figure 2.33).

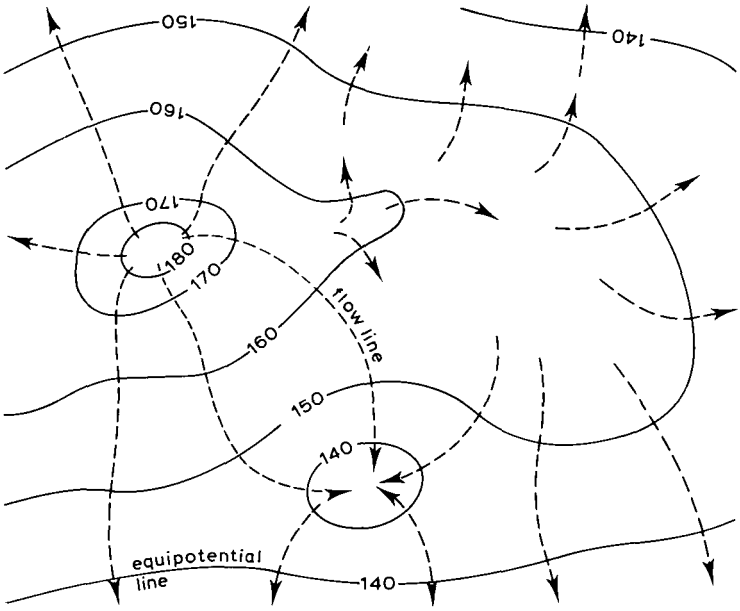

Figure 2.32 A watertable-contour map showing a local mound and a depression in the watertable and the direction of groundwater flow

In areas where a stream is losing water to the underground, the watertable contours have bends in downstream direction. At places where the contours are at right angles to the stream, groundwater is flowing neither towards nor from the stream but down the general slope of the watertable. In areas where groundwater is flowing into the stream, the watertable contours have bends in upstream direction.

The bends in watertable contours near streams and drainage channels may have different shapes due to differences in the resistance to radial flow; the longer and narrower the bends, the higher this flow resistance is. Obviously, to determine the shape of the bends, water-level readings in several observation wells in the near vicinity of the stream are required, as outlined earlier.

Ernst (1962) has presented an equation to estimate the value of the radial resistance, which is the resistance that groundwater has to overcome while flowing into a stream or drainage channel because of the contraction of the flow lines in the vicinity of the stream.

For a proper interpretation of a watertable-contour map, one has to consider not only the topography, natural drainage pattern, and local recharge and discharge patterns, but also the subsurface geology. More specifically, one should know the spatial distribution of permeable and less permeable layers below the watertable. For instance, a clay lens impedes the downward flow of excess irrigation water or, if the area is not irrigated, the downward flow of excess rainfall. A groundwater mound will form above such a horizontal barrier (Figure 2.34).

The surface of the first effective impermeable layer below the watertable can be undulating and, when viewed over greater distances, it can be dipping. At some places, the impermeable layer may rise to or close to the land surface. If such a ridge of tight

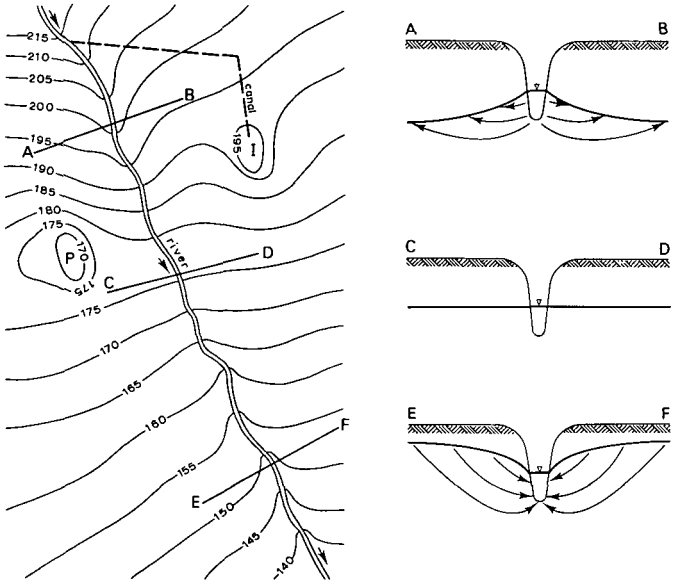

Figure 2.33 A watertable-contour map showing a stream that is losing water in its upstream part and gaining water in its downstream part; in its middle part, it is neither gaining nor losing water. I: Irrigation causes local mound in the watertable; P: Pumping causes local depression in the watertable

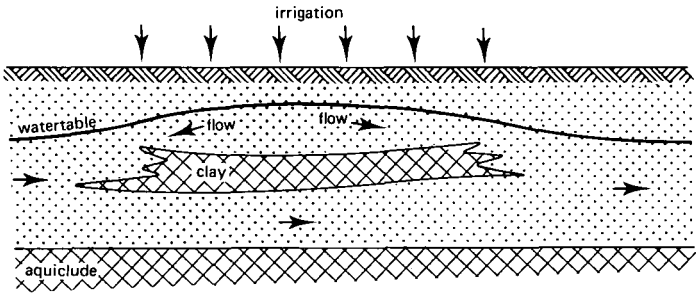

Figure 2.34 A clay lens under an irrigated area impedes the downward flow of excess irrigation water

clay occurs at right angles to the general groundwater flow, the natural drainage will be blocked (Figure 2.35).

Watertable-contour maps are graphic representations of the hydraulic gradient of the watertable. The velocity of groundwater flow (v) varies directly with the hydraulic gradient (dh/dx) and, at constant flow velocity, the gradient is inversely related to the hydraulic conductivity (K), or $v = -K(dh/dx)$ (Darcy's law). This is a fundamental law governing the interpretation of hydraulic gradients of watertables. (For a further discussion of this law, see Chapter 7.) Suppose the flow velocity in two cross-sections of equal depth and width is the same, but one cross-section shows a greater hydraulic gradient than the other, then its hydraulic conductivity must be lower. A steepening of the hydraulic gradient may thus be found at the boundary of fine-textured and coarse-textured material (Figure 2.36A), or at a fault where the thickness of the water-bearing layers changes abruptly (Figure 2.36B).

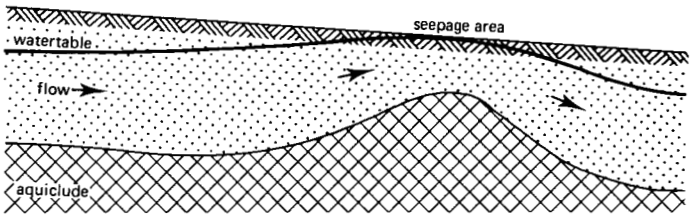

Figure 2.35 Effect of an impermeable barrier on the watertable

Characteristic groundwater conditions can be found in river plains. In the humid zones, such plains usually have the profile shown in Figure 2.37A. Silted-up former stream channels are common in these plains, and if their sand body is in direct contact with an underlying coarse sand and gravel layer whose groundwater is under pressure, they form important leaks in the low-permeable covering layer. Figure 2.37B shows the distribution of the piezometric head/watertable elevation at different depths in a row of piezometers perpendicular to the buried stream channel.

Depth-to-Watertable Maps

From our discussions so far, it will be clear that a variety of factors must be considered if one is to interpret a depth-to-groundwater or isobath map properly. Shallow watertables may occur temporarily, which means that the natural groundwater runoff cannot cope with an incidental precipitation surplus or irrigation percolation. They may occur (almost) permanently because the inflow of groundwater exceeds the outflow, or groundwater outflow is lacking as in topographic depressions. The depth and shape of the first impermeable layer below the watertable strongly affect the height of the watertable. To explain differences and variations in the depth to watertable, one has to consider topography, surface and subsurface geology, climate, direction and rate of groundwater flow, land use, vegetation, irrigation, and the abstraction of groundwater by wells.

Watertable-Fluctuation Maps

The watertable in topographic highs is usually deep, whereas in topographic lows it is shallow. This means that on topographic highs there is sufficient space for the watertable to change. This space is lacking in topographic lows where the watertable

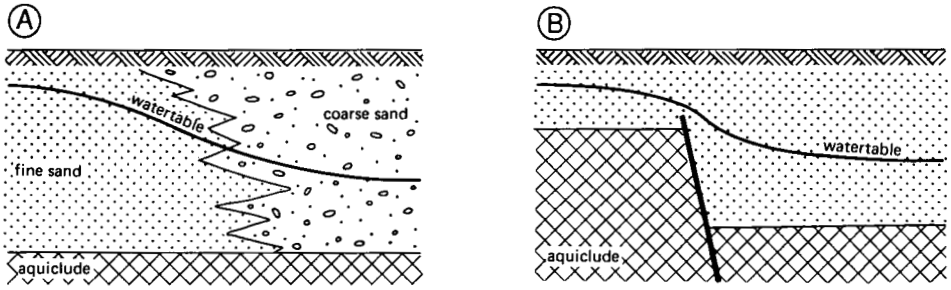

Figure 2.36 Examples of the effect on the hydraulic gradient
A: Of permeability; B: Of bed thickness

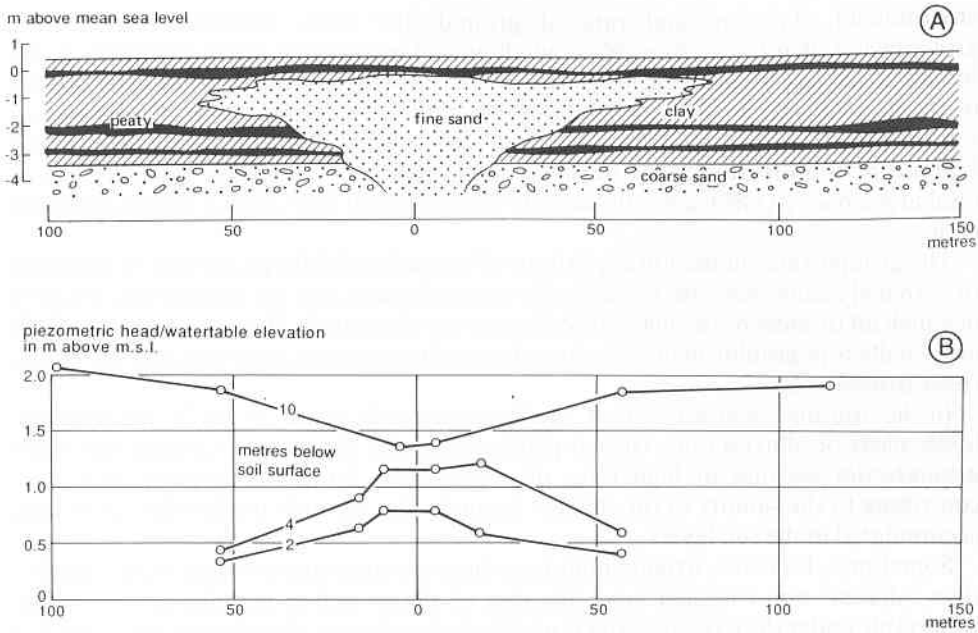

Figure 2.37 Characteristic groundwater condition in river plain
A: Cross-section of a Holocene silted-up former stream channel of the River Waal (after Verbraeck and de Ridder 1962); B: Distribution of hydraulic heads at 2, 4, and 10 m below ground surface along a line perpendicular to a sand-filled former river channel as shown in Figure 2.37A. The water level in the nearby River Waal at the time of observation was 3.70 m above mean sea level (after Colenbrander 1962)

is often close to the surface. Watertable fluctuations are therefore closely related to depth to groundwater.

Another factor to consider in interpreting watertable-fluctuation maps is the drainable pore space of the soil. The change in watertable in fine-textured soils will differ from that in coarse-textured soils, for the same recharge or discharge.

Head-Differences Maps

The difference in hydraulic head between the shallow and the deep groundwater is directly related to the hydraulic resistance of the low-permeable layer(s). Because such layers are seldom homogeneous and equally thick throughout an area, the hydraulic resistance of these layers varies from one place to another. Consequently, the head difference between shallow and deep groundwater varies. Local 'leaks' in low-permeable layers may result in anomalous differences in hydraulic heads, as was demonstrated in Figure 2.37B. The hydraulic resistance is especially of interest when one is defining upward seepage or natural drainage (Chapters 9 and 16) or the possibilities for tubewell drainage (Chapters 10 and 22).

Groundwater-Quality Maps

Spatial variations in groundwater quality are closely related to topography, geological

environment, direction and rate of groundwater flow, residence time of the groundwater, depth to watertable, and climate. Topographic highs, especially in the humid zones, are areas of recharge if their permeability is fair to good. The quality of the groundwater in such areas almost resembles that of rainwater. On its way to topographic lows (areas of discharge), the groundwater becomes more mineralized because of the dissolution of minerals. Although the water may be still fresh in discharge areas, its electrical conductivity can be several times higher than in recharge areas.

The groundwater in the lower portions of coastal and delta plains may be brackish to extremely salty, because of sea-water encroachment and the marine environment in which all or most of the mass of sediments was deposited. Their upper parts, which are usually topographic highs, are nowadays recharge areas and consequently contain fresh groundwater.

In the arid and semi-arid zones, shallow watertable areas, as can be found in the lower parts of alluvial fans, coastal plains, and delta plains, may contain very salty groundwater because of high rates of evaporation. Irrigation in such areas may contribute to the salinity of the shallow groundwater through the dissolution of salts accumulated in the soil layers.

Sometimes, however, irrigated land can have groundwater of much better quality than adjacent non-irrigated land. Because of the irrigation percolation losses, the watertable under the irrigated land is usually higher than in the adjacent non-irrigated land. Consequently, there is a continuous transport of salt-bearing groundwater from the irrigated to the non-irrigated land. This causes the watertable in the non-irrigated land to rise to close to the surface, where evapo(transpi)ration further contributes to the salinization of groundwater and soil (Chapter 15).

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