Studies and reports in hydrology 7

Ground-water studies

An international guide for research and practice

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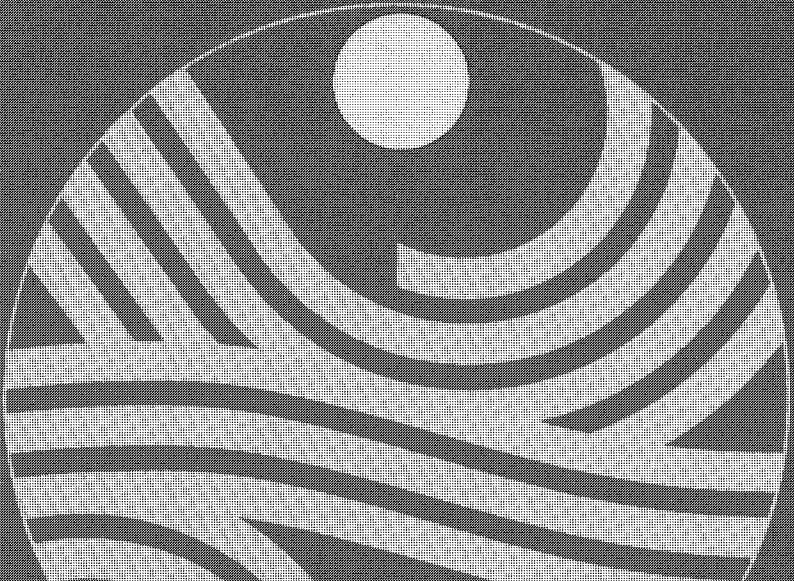

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7 Observations and instruments

7.1 Location of observation wells

The number of observation wells necessary to study the ground-water régime, and their areal distribution, is dependent on the special features of the area to be investigated and the type of study planned.

For example, if a thorough study of the natural occurrences and interrelationships of ground and surface water is intended for a basin (having an area of 100 km^2 or less) the annual and seasonal ground-water balances as well as total water balances must be determined. Basins of this type may be considered as complex water-balance laboratories under natural conditions. Observations made in such basins allow the determination and study of various natural processes that govern the quality and distribution of waters of various types.

For a basin having an area of up to 1,000 km^2 only the principal elements of the water balance are usually studied. These are precipitation, run-off, maximum snow storage and, in some cases, the ground-water régime and other elements.

A general discussion of criteria for designing a network of observation wells is given later in this chapter in Section 7.1.2. Network planning was the subject of symposia at Quebec, Canada (WMO-IASH, 1965) and Berne, Switzerland (IASH, 1968).

The special features of investigations and distribution of observation-well networks for the following types of ground-water régimes or aquifer systems will be considered:

1. Platform areas: (a) homogeneous areas; (b) heterogeneous areas; (c) complex areas.
2. Mountain-folded areas: (a) areas with strip-like occurrences of permeable rocks; (b) folded areas.
3. Permafrost areas.
4. Submontane terrains and valleys in arid regions: (a) pebbled (coarsest) part of the alluvial fan; (b) the middle part of the alluvial fan; (c) the subzone of unconfined ground-water discharge; (d) subzone of resubmergence and evaporation of ground water; (e) subzone of alluvial ground water.
5. Alluvial and outwash plains (zone of excess moisture).
6. Karst areas.

This classification of ground-water régimes is arbitrary and by no means comprehensive; however, it allows the development of guides for the planning and programming of observation well networks.

7.1.1 *Environments affecting the design of well networks*

7.1.1.1 **Platform areas**

The principal feature of the geological structure in platform areas is the horizontal attitude of sedimentary rocks which overlie the basement rocks and the general absence of tectonic disturbances. Ground water in the sedimentary rocks occurs commonly in permeable sands drained by streams and recharged primarily by infiltration of precipitation.

Homogeneous conditions. Under homogeneous conditions the sedimentary rocks of the area constitute commonly a single unconfined, or water-table, aquifer. Temporarily perched ground water may occur on clay lenses in the zone of aeration (Fig. 7.1.1.1a, I.1). Streams drain the water-table aquifers completely or partially depending on the depth of their downcutting. The aquifers are recharged mainly by local precipitation and the topographic divide of the basin generally coincides with the ground-water divide.

The base flow, Y_b, from a basin of this type may be determined by separating a streamflow hydrograph obtained at the basin outlet. If the under-channel run-off at the outlet is less than 5 per cent of the base flow, it need not be considered in the stream basin balance. By analysis of water-table fluctuations the value of ground-water recharge from above, $W\Delta t$, can be determined, and knowing the change in ground-water storage, $\mu\Delta H$, the value of ground-water flow, Q_{II}, from the drainage basin area may be determined from the following equation:

$$Y_b = Q_{II} = \sum_{T} W_{\text{w.av.}}\, \Delta t - \sum_{T} (\mu\Delta H)_{\text{w.av.}} \qquad 7.1(1)$$

where summations represent the products of the weighted average values of recharge, $W_{\text{w.av.}}\Delta t$, and the weighted averages values, $(\mu\Delta H)_{\text{w.av.}}$, of the change in ground-water storage.

The unconfined ground-water balance for the basin is composed of the weighted average (in respect to area) values of infiltration of precipitation, $I_{\text{w.av.}}\,\Delta\tau$; values of evaporation from the water table, $U_{\text{w.av.}}\Delta V$; and values of the differences between the inflow Q_1 and outflow Q_2 of unconfined ground water $[(Q_1 - Q_2)/F]_{\text{w.av.}}\Delta t$. Values for these elements are calculated from data obtained from ground-water régime observations made on calculation (balance) plots. Thus the unconfined ground-water balance is determined by use of the equation:

$$\sum_{T} (\mu\Delta H)_{\text{w.av.}} = \sum_{T} \left(\frac{Q_1 - Q_2}{F}\right)_{\text{w.av.}} \Delta t + \sum_{T} I_{\text{w.av.}}\, \Delta\tau - \sum_{T} U_{\text{w.av.}}\, \Delta V \qquad 7.1(2)$$

The total water balance for the basin is determined from the formula:

$$(\mu\Delta H)_{\text{av}} + W_1 + W_2 = X - Z + K_1 - Y_{2,\text{ov}} - Q_{II} \qquad 7.1(3)$$

where $Y_{2,\text{ov}} = Y_2 - Q_{II}$, is the overland flow from the basin.

Heterogeneous conditions. Under heterogeneous conditions there are two or more discrete aquifers in the basin in addition to any temporarily perched water.

The base flow is composed of ground-water flow from the first and second aquifers (Fig. 7.1.1.1a, I.2). As the ground water in the second aquifer results from leakage from the first aquifer the total base flow may be estimated either by stream hydrograph separation (as in the case of homogeneous conditions) or by use of Equation 7.1(1).

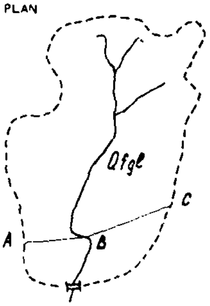
The unconfined ground-water balance differs from that for homogeneous conditions because of the vertical water exchange between aquifers as indicated by the differences in hydraulic head in the first and second aquifers. The amount of this exchange between aquifers is dependent on the hydraulic conductivity of the intervening semi-pervious bed

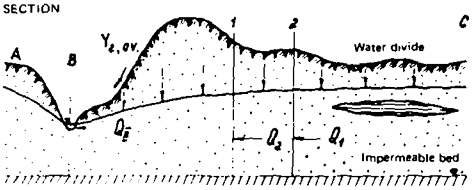
FIG. 7.1.1.1a. Typical diagrams of structure in representative basins

I. PLATFORM AREAS. 1.1. Homogeneous conditions

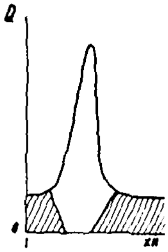
Geological and hydrogeological conditions

Stream hydrograph separation





A single main aquifer of temporary perched water.
All ground-water flow drained by the stream.



Pattern of base flow Y_b (mm/year)

$$Y_b = Q_{II} = \sum_t W_{w.av.}\Delta t - \sum_t (\mu\Delta H)_{w.av.}$$

Pattern of the unconfined ground-water balance for a basin with an area A (mm/year)

$$\sum_t (\mu\Delta H)_{w.av.} = \sum_t \left(\frac{Q_1 - Q_2}{F}\right)_{w.av.} \Delta t + \sum_t I_{w.av.}\Delta\tau - \sum_t U_{w.av.}\Delta V$$

Pattern of the total water balance for a basin with an area A (mm/year)

$$(\mu\Delta H)_{av.} + W_1 + W_2 = X - Z + K_1 - Y_{2,ov.} - Q_{II}$$

where: $$Y_{2,ov.} = Y_2 - Q_{II}$$

Type of base flow régime: Primarily derived from bank storage.

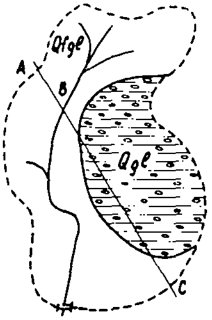
FIG. 7.1.1.1a. Typical diagrams of structure in representative basins (cont.)

I. PLATFORM AREAS. I.2. Heterogeneous conditions

Geological and hydrogeological conditions

Stream hydrograph separation

PLAN



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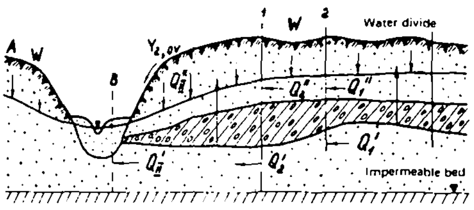
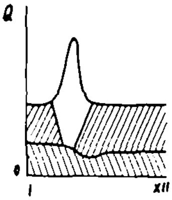


Two aquifers apart from temporary perched water. Ground-water flow from the lower aquifer is equal to the amount of leakage from the upper aquifer.



Pattern of base flow Y_b (mm/year)

$$Y_b = Q'_{II} + Q''_{II} = \sum_t W_{w.av.}\Delta t - \sum_t (\mu\Delta H)_{w.av.}$$

Pattern of the unconfined ground-water balance for a basin with an area A (mm/year)

$$\sum_t (\mu\Delta H)_{w.av.} = \sum_t \left(\frac{Q_1 - Q_2}{F}\right)_{g.w.av.} \Delta t - \sum_t U_{w.av.}\Delta V - \sum_t \left(\frac{Q\text{asc.}}{F}\right)_{w.av.} \Delta t + \sum_t I_{w.av.}\Delta\tau$$

Pattern of the total water balance for a basin with an area A (mm/year)

$$(\mu\Delta H)_{av.} + W_1 + W_2 = X - Z + K_1 - Y_{2,ov.} - Q''_{II} - Q'_{II}$$

where: $Y_{2,ov.} = Y_2 - Q''_{II} - Q'_{II}$

Type of base flow régime: Primarily from bank storage.

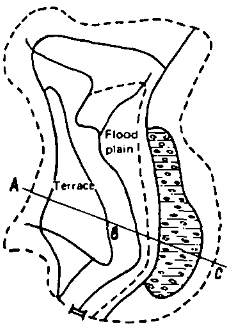
FIG. 7.1.1.1a. Typical diagrams of structure in representative basins (cont.)

I. PLATFORM AREAS. I.3. Complex conditions

Geological and hydrogeological conditions

Stream hydrograph separation

PLAN



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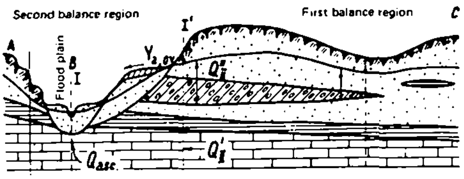
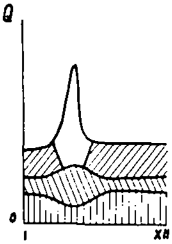


Three aquifers apart from temporary perched water.
A stream drains the confined aquifer the recharge area of which does not coincide with the basin.



Pattern of base flow Y_b *(mm/year)*

$$Y_b = Q_{asc.} - Q_{II}' + Q_{II}'' - Z_{I ter.} - Z_{fl.pl.}$$

$$\text{At } X \neq 0;\ \Delta H \neq 0,\ Q_{II} = Q_{II}' + Q_{II}'' + X_{I-I'} - Z_{I-I'} - Y_{2,ov.} - (\mu \Delta H)_{av.\ I-I'}$$

$$Q_{asc.} = Y_b - Q_{II}$$

(Proved by hydrodynamic calculation.)

Pattern of the unconfined ground-water balance for a basin with an area A (mm/year)

The same as before, for unconfined ground water, but for separate regions with conditions which are similar to diagrams I.1 and I.2. The ground-water balance is made individually for separate regions.

Pattern of the total water balance for a basin with an area A (mm/year)

The same as before, for the part of basin that is to the right of section I-I'. For all the basin:

$$(\mu \Delta H)_{av.} + W_1 + W_2 = X - Z + K_1 - Y_{2,ov.} - Q_I' - Q_{II}'' =$$

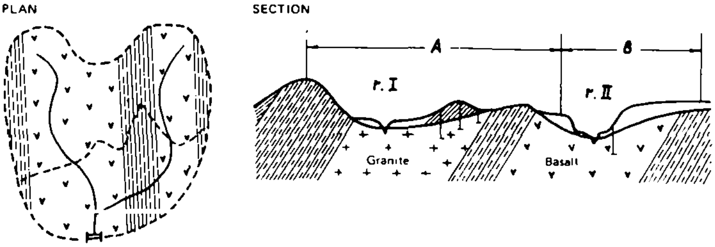
$$= X_{I-I'} + Z_{I-I'} + Y_{2,ov.I-I'} + (\mu \Delta H)_{av.I-I'} + Q_{asc}.$$

Type of base flow régime: Cascading springs and seeps and bank storage, and artesian discharge.

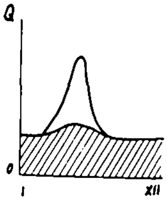
FIG. 7.1.1.1a. Typical diagrams of structure in representative basins. (cont)

II. MOUNTAIN-FOLDED AREAS. II.1. Areas with a strip-like occurrence of permeable rocks

Geological and hydrogeological conditions



Stream hydrograph separation



Pattern of base flow Y_b (mm/year)

$$Y_b = Q_{II} = \sum_t W_{w.av.} \Delta t - \sum_t (\mu \Delta H)_{w.av.}$$

Pattern of the unconfined ground-water balance for a basin with an area A (mm/year)

The ground-water balance is made for individual tributary basin-blocks of hydrogeological structure.

$$\sum_t (\mu \Delta H)_{w.av.} - \sum_t \left(\frac{Q_1 - Q_2}{F} \right)_{w.av.} \Delta t + \sum_t W_{w.av.} \Delta t$$

Pattern of the total water balance for a basin with an area A (mm/year)

$$(\mu \Delta H)_{av.} + W_1 = X - Z + K_1 - Y_2$$

$$W_2 = 0$$

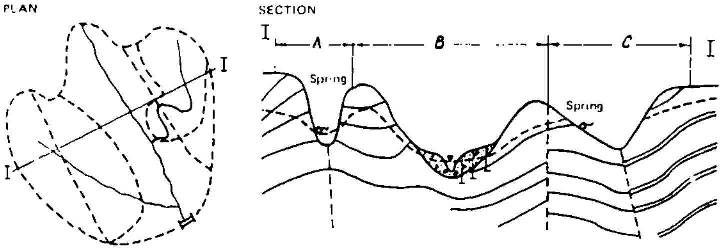
$$Y_2 = Y_{2,ov.} + Y_b \text{ (from separation of streamflow)}$$

Type of base flow régime: Cascading springs and seeps and bank storage.

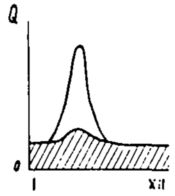
FIG. 7.1.1.1a. Typical diagrams of structure in representative basins (cont.)

II. MOUNTAIN-FOLDED AREAS. II.2. Folded areas

Geological and hydrogeological conditions



Stream hydrograph separation



Pattern of base flow Y_b (mm/year)

$$Y_b = \sum_t W_{\text{w.av.}}\Delta t = X - Z - K_1 - Y_{2,\text{ov.}} - W_1$$

$$W_2 \approx 0$$

$$\mu \Delta H \approx 0$$

$$\mu \approx 0$$

Pattern of the unconfined ground-water balance for a basin with an area A (mm/year)

The ground-water balance is made for individual basins.

$$\sum_t (\mu \Delta H)_{\text{w.av.}} = X - Z + K_1 - W_1 - Y_{2,\text{ov.}} - Y_b = 0$$

$$W_1 = 0$$

Pattern of the total water balance for a basin with an area A (mm/year)

The water balance is made for individual tributary basins.

$$(\mu \Delta H)_{\text{av.}} + W_1 = X - Z + K_1 - Y_{2,\text{ov.}} - Y_b$$

Type of base flow régime: Cascading springs.

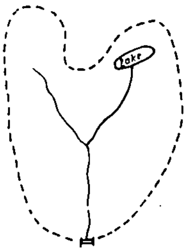
FIG. 7.1.1.1a. Typical diagrams of structure in representative basins (cont.)

III. PERMAFROST AREAS

Geological and hydrogeological conditions

Stream hydrograph separation

PLAN



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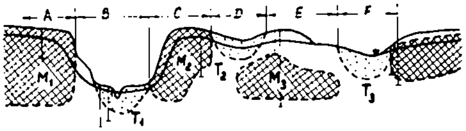
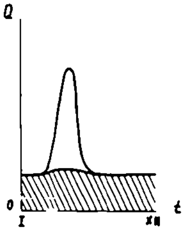


A. Key plot at the junction of the seasonal thawing layer (M_1) with perennially frozen rocks.
B. In a talik under the stream.
C. The same as A.
D. Closed talik.
E. Region of discontinuous frozen ground.
F. Talik under a lake.



Pattern of base flow Y_b (mm/year)

$$Y_b = Q_{II}$$

Pattern of the unconfined ground-water balance for a basin with an area A (mm/year)

A, C. (see section). In layer of seasonal thawing of permafrost rocks:

$$\sum_t (\mu \Delta H)_{\text{w.av.}} = \sum_t \left(\frac{Q_1 - Q_2}{F}\right)_{\text{w.av.}} \Delta t + W\Delta t$$

$$W\Delta t = X - Z + K_1 - W_2 - W_1 - Y_{2,\text{ov.}}$$

B. In talik near a stream.

E. In region of discontinuous frozen ground:

$$w\Delta t = X - Z + K_1 - W_1 - W_2 - Y_{2,\text{ov.}}$$

Pattern of the total water balance for a basin with an area A (mm/year)

A, C: $(\mu\Delta H)_{\text{av.}} + W_1 + W_2 = X - Z + K_1 - Y_{2,\text{ov.}} - Q_{II}$

B: $(\mu\Delta H)_{\text{av.}} + W_1 + W_2 = X - Z + K_1 - Y_{1,\text{ov.}} - Y_{2,\text{ov.}} - Q_I - Q_{II}$

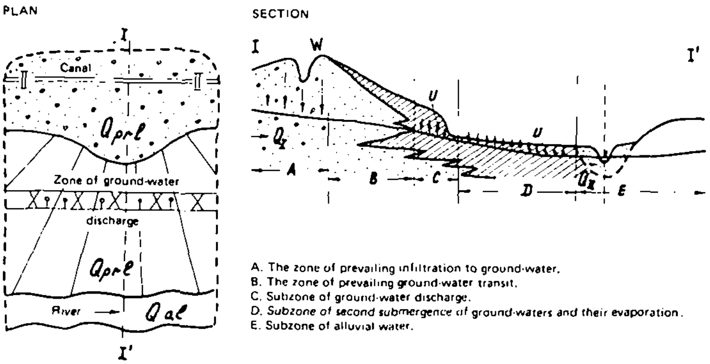
E: $(\mu\Delta H)_{\text{av.}} + W_1 + W_2 = X - Z + K_1 - Y_{2,\text{ov.}} - Q_{II}$

Type of base flow régime: Cascading springs and seeps; bank storage with added flow from melting of previously frozen ground-water discharge.

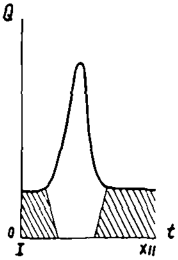
FIG. 7.1.1.1a. Typical diagrams of structure in representative basins (cont.)

IV. ZONE OF SUBMONTANE TERRAINS AND VALLEYS (ARID ZONE)

Geological and hydrogeological conditions | *Stream hydrograph separation*



A. The zone of prevailing infiltration to ground-water.
B. The zone of prevailing ground-water transit.
C. Subzone of ground-water discharge.
D. Subzone of second submergence of ground-waters and their evaporation.
E. Subzone of alluvial water.



Pattern of base flow Y_b *(mm/year)*

$Y_b = Q_{II}$

Pattern of the unconfined ground-water balance for a basin with an area A (mm/year)

A:
$$\sum_t(\mu\Delta H)_{\text{w.av.}} = \sum_t\left(\frac{Q_1-Q_2}{F}\right)_{\text{w.av.}}\Delta t + \sum_t I_{\text{w.av.}}\Delta\tau$$

B:
$$\sum_t(\mu\Delta H)_{\text{w.av.}} = \sum_t\left(\frac{Q_1-Q_2}{F}\right)_{\text{w.av.}}\Delta t$$

C:
$$\sum_t(\mu\Delta H) = \sum_t\left(\frac{Q_1-Q_2}{F}\right)\Delta t + U\Delta V$$

D:
$$\sum_t(\mu\Delta H) = \sum_t\left(\frac{Q_1-Q_2}{F}\right)\Delta t + \sum_t I\Delta\tau - \sum_t U\Delta V$$

E:
$$\sum_t(\mu\Delta H) = \sum_t\left(\frac{Q_1-Q_2}{F}\right)\Delta t + \sum_t I\Delta\tau$$

Pattern of the total water balance for a basin with an area A (mm/year)

A, B, C: $(\mu\Delta H)_{\text{av.}} + W_1 + W_2 = X - Z + K_1 - Y_{\text{ov.}} - Q_I - Q_{II}$

D: $(\mu\Delta H)_{\text{av.}} + W_1 + W_2 = X - Z - Y_{\text{ov.}} + Q_I - Q_b - Q_{II}$

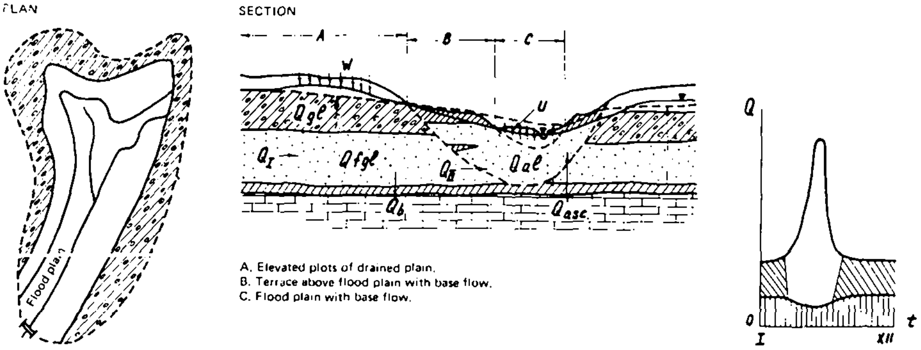
E: $(\mu\Delta H)_{\text{av.}} + W_1 + W_2 = X - Z + Q_I - Q_{II}$

Type of base flow régime: Commonly derived from bank storage.

FIG. 7.1.1.1a. Typical diagrams of structure in representative basins (cont.)

V. AREAS OF ALLUVIAL AND OUTWASH PLAINS (ZONE OF EXCESSIVE MOISTENING)

Geological and hydrogeological conditions *Stream hydrograph separation*



A. Elevated plots of drained plain.
B. Terrace above flood plain with base flow.
C. Flood plain with base flow.

Pattern of base flow Y_b (mm/year)

$$Y_b = Q_{II} + Q_{asc.}$$

Pattern of the unconfined ground-water balance for a basin with an area A in zones (mm/year)

A. In drained plots:

$$\sum_t(\mu\Delta H) = \sum_t\left(\frac{Q_1-Q_2}{F}\right)\Delta t + \sum_t I\Delta\tau - \sum_t U\Delta V - \frac{Q_b}{F}\Delta t$$

B. Over flood plain:

$$\sum_t(\mu\Delta H) = \sum_t\left(\frac{Q_1-Q_2}{F}\right)\Delta t + \frac{Q_{asc.}}{F}\Delta t - U\Delta V$$

C. Flood plain:

$$\sum_t(\mu\Delta H) = \sum_t\left(\frac{Q_1-Q_2}{F}\right)\Delta t + \frac{Q_{asc.}}{F}\Delta t + I\Delta\tau - U\Delta V$$

Pattern of the total water balance for a basin with an area A in zones (mm/year)

A. In drained plots:

$$(\mu\Delta H)_{av.} + W_1 + W_2 = X - Z + K_1 - Y_{2,ov.} + Q_I - Q_{II}$$

B. Over flood plain:

$$(\mu\Delta H)_{av.} + W_1 + W_2 = X - Z + K_1 - Y_{ov.} + Q_I + Q_b - Q_{II}$$

C. Flood plain:

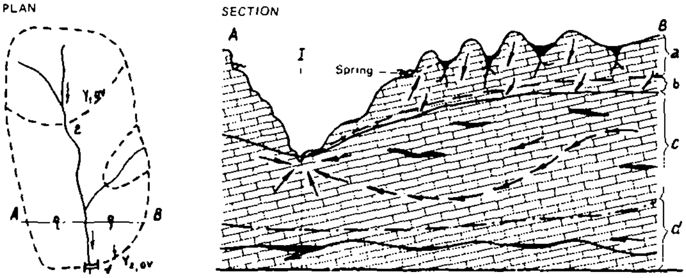
$$(\mu\Delta H)_{av.} + W_1 + W_2 = X - Z + K_1 - Y_{ov.} + Q_I + Q_{asc.} - Q_{II}$$

Type of base flow régime: Cascading springs and seeps; bank storage and artesian discharge; occasional stream loss to ground water.

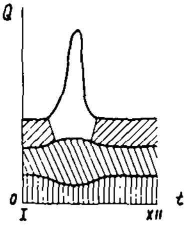
FIG. 7.1.1.1a. Typical diagrams of structure in representative basins (cont.)

VI. KARST AREAS

Geological and hydrogeological conditions



Stream hydrograph separation



Pattern of base flow Y_b (mm/year)

$$Y_b = \sum_t w_{w.av.}\Delta t + Q_{asc.} - \sum_t (\mu\Delta H)_{w.av.}$$

Pattern of the unconfined ground-water balance for a basin with an area A (mm/year)

$$\sum_t (\mu\Delta H)_{av.} = X - Z + K_1 - W_1 - Y_{2,ov.} - Y_b + Q_{asc.} + Y_{1,ov.}$$

Pattern of the total water balance for a basin with an area A (mm/year)

$$(\mu\Delta H)_{av.} + W_1 = X - Z - K_1 - Y_{1,ov.} - Y_{2,ov.} + Q_{asc.} - Y_b$$

Type of base flow régime: Cascading springs and seeps; bank storage and artesian discharge.

and the hydraulic gradient. The total water balance also includes the base-flow contribution from the lower aquifer.

Complex conditions. Basins exhibiting complex conditions are characterized by multi-layered water-bearing rocks and by the drainage of two or more surficial aquifers, as well as some vertical drainage of confined ground water from permeable basement rocks, into the streams (Fig. 7.1.1.1a, I.3). The water discharged from basement rocks (ascending water) may have come from recharge areas distant from the stream.

In the example shown on Figure 7.1.1.1a, I.3, the aquifer underlain by moraine crops out at the upper terrace and the aquifer underlying the moraine is drained by the first terrrace above the flood plain; this water, in turn, discharges to the flood-plain terrace and eventually enters the stream.

Under such conditions, the base-flow in section *I-I′* will be composed of the discharge of ascending ground water Q_{asc} drained by the stream and the discharge of unconfined ground-water flow Q_g from alluvial terraces. The unconfined ground-water flow Q_{II} may be found by using the data on ground-water outflow from interstratal waters (on the diagram: from above-moraine and under-moraine aquifers, Q''_{II} and Q'_{II} respectively in section *I-I′* and data on the evaporation from the surface of the terrace and the flood plain if the overland flow and precipitation are absent (if the change in the level ΔH is equal to 0).

With precipitation X over the valley area and water-level fluctuation $\Delta H \neq 0$ over the time interval Δt, the base flow may be determined from the formula:

$$Q_{II} = Q'_{II} + Q''_{II} + X_{I-I'} - Z_{I-I'} - Y_{2,\mathrm{ov}} + (\mu \Delta H)_{\mathrm{av},I-I'} \qquad 7.1(4)$$

In this equation the precipitation $X_{\mathrm{I-I}'}$ evapotranspiration $Z_{\mathrm{I-I}'}$ and overland flow $Y_{2,\mathrm{ov}}$ in the valley, expressed in millimetres of a water layer, relate to the entire drainage basin.

The amount of confined ground water (Q_{asc}) drained from the lower aquifer may be determined from the stream hydrograph if Q_{II} is known or from hydraulic relationships if the permeability of the stream bed and the piezometric gradient of the vertical seepage are known.

The unconfined ground-water balance is found from the following relationship (differentially for regions of the drainage basin):

$$\sum_T (\mu \Delta H)_{\mathrm{w.av.}} = \sum_T \left(\frac{Q_1 - Q_2}{F}\right)_{\mathrm{w.av.}} \Delta t - \sum_T U_{\mathrm{w.av.}}\, \Delta V -$$

$$- \sum_T \left(\frac{Q_{asc}}{F}\right)_{\mathrm{w.av.}} \Delta t + \sum_T I_{\mathrm{w.av.}}\, \Delta \tau \qquad 7.1(5)$$

To apply the unconfined ground-water balance methods to the study of a basin having an *area of less than about 100 km*^{2}, requires the following:

1. On each type of microlandscape,[1] no less than 3 to 5 key observation plots differing in depths to the water table should be established. On each plot a line of no less than 3 wells should be installed along the direction of ground-water flow. These wells will furnish data for calculating ground-water recharge from infiltration by use of finite-difference methods. The distance between wells should usually be several hundred metres (200-300 metres is better if the water-level difference between wells is more than 0.20 – 0.30 m). For basins with a complex structure a greater number of wells may be needed. Conversely, if the basin has a very simple structure, one line of observation wells installed near the outlet is often sufficient.
2. Data from observation wells on lines near streams may be used for calculating the

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[1] The microlandscape type is a small area that is uniform as to topography, soils, vegetation, geological structure, moisture conditions on the land surface and ground-water flow.

hydraulic diffusivity of the aquifer, T/S. The distance from the river of any well used for this purpose should not be more than $2\sqrt{(Tt/S)}$, where t is the time period for the abnormal fluctuation of river level that is monitored.

3. The hydraulic conductivity for each type of water-bearing rock should be determined by means of pumping tests. If the hydraulic diffusivity has been determined for thick aquifers from water-level fluctuations, the hydraulic conductivity need not be determined.

If the representative basin being studied is *larger than about 100 km²* it is usually not practical from an economic standpoint or necessary from a hydrologic standpoint to install calculation plots on every microlandscape in the basin. Plots need be installed only on principal or typical microlandscapes, the selection of which are dependent on the professional judgement of the investigators. Single observation wells may be installed to supply additional data between the plots.

After two or three years of observations of the ground-water régime it is usually possible to correlate water levels in various wells and reduce the number of observations.

When an observation well is drilled more than 1 km from an existing one, soil samples should be taken from the zone of aeration above the capillary fringe for determination of

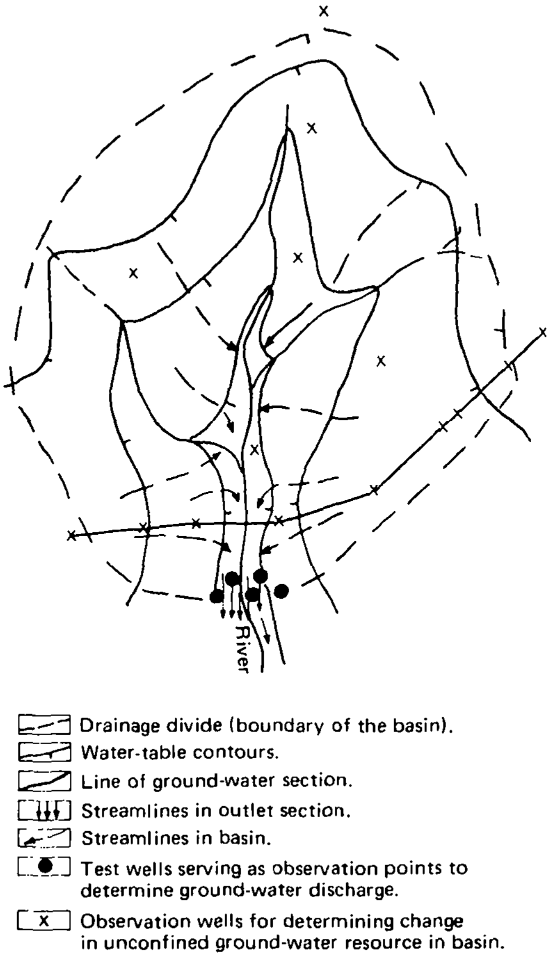


FIG. 7.1.1.1b. Arrangement of lines of observation wells in a river basin.

maximum and minimum moisture content and the volume moisture content V_0. The values of these determinations are then used in water-balance calculations of changes in unconfined ground-water storage, particularly to determine specific yield.

Representative basins having *areas of up to 1,000 km²* are studied according to a reduced programme. Observations are made of the principal elements of the water balance: streamflow, precipitation, maximum snow storage and in some cases the ground-water régime and some other elements of the balance.

For the IHD programme, observation wells are installed in representative large basins according to the principles for small basins but not at the same density—numbers, locations and types of wells being dependent on basin conditions. Most wells are intended to supply data on seasonal and long-term water-level fluctuations and on the relationships between the ground-water and surface-water régimes.

Under proper hydrogeological conditions most wells can be located along a line from divide to divide crossing both slopes of the river basin near the outlet (Fig. 7.1.1.1b). The typical features of the basin's ground-water régime can be elucidated as a result of these observations. If this requirement is not realized additional wells must be constructed as needed.

7.1.1.2 Mountain-folded areas

In general, the characteristic features of representative basins in these areas are the sharp ruggedness of the relief, non-uniformity of rocks as to their composition and water-bearing properties, occurrence of ground water in fractures and wide distribution of spring discharges.

An area underlain by rocks that differ widely in permeability is typical of these basins. The rocks have been tilted and eroded so that their outcrop pattern is such that strips of more permeable rocks alternate with less permeable rocks (Fig. 7.1.1.1a, II.1).

The ground-water balance for each unit must be studied. The base flow, or ground-water flow, from the area may be determined by stream-hydrograph separation or by measuring the ground-water increment by making discharge measurements during conditions of low flow.

If a water-bearing crust of weathered rock is present, data obtained from a limited network of shallow observation wells may be used to estimate variations in ground-water storage ($\mu\Delta H$) in this material. The value of ground-water recharge $(W\Delta t)_{av}$ in various parts of the basin may be estimated from data obtained by observations of precipitation, evapotranspiration and overland flow. The average long-term value of this recharge is equal to the mean base flow separated from stream hydrographs.

In compiling a total water balance for an area of this type the change in moisture reserves in the zone of aeration may be neglected ($W_2 = 0$) because of the extremely small moisture storage capacity of this zone. Typical springs should be instrumented to allow systematic observations to assist in hydrograph separation and to determine total spring discharge values. Occasional records may be obtained from other springs to supplement these systematic records.

Folded and faulted rock strata that have undergone deformation sufficient to hamper their separation into uniform blocks or strips may also be typical of mountain basins (Fig. 7.1.1.1a, II.2). Slope talus, deep erosional down-cutting and numerous springs are characteristic of this mountainous relief.

Due to the drainage characteristics of the rocks and their low moisture retention capacity the change in ground-water storage ($\mu\Delta H = 0$) may be neglected. The base flow then is determined from the formula:

$$Y_b = \sum_T W_{w.av.}\,\Delta t = X - Z + K_1 - Y_{2,ov} - W_1 \qquad 7.1(6)$$

The unconfined ground-water balance may be determined by the use of data on tributary basins:

$$\sum_{T} (\mu \Delta H)_{\text{w.av.}} = X - Z + K_1 - W_1 - Y_{2,\text{ov}} - Y_b = 0 \qquad 7.1(7)$$

The total water balance is also determined by use of data from tributary basins. In principle, this balance does not differ from the unconfined ground-water balance because the change in ground-water storage, $\mu \Delta H$, and the moisture storage change in the zone of aeration may be neglected for representative basins of this type.

To understand the ground-water régime in areas of this type it is necessary to apportion the total ground-water yield of the basin, as determined at its outlets, to the yields of the various tributary streams. The sub-basins may have widely varying ground-water yields because of permeability differences related to petrographic and genetic variations in rock types. The manner in which the division of total basin yield into its sub-basin components can be accomplished is dependent upon the geologic and geomorphic characteristics of the area.

Consider an area characterized by alternating strip-like outcrops of relatively permeable and impermeable rocks (Fig. 7.1.1.1a, II.1) in which the stream courses are coincident with the strips of permeable rock. Measuring sections need only be installed along the streams to determine balances for specified reaches. Particular attention should be given to determining whether or not the ground-water recharge area of the basin or sub-basin coincides with its surface drainage area.

If the tributaries or the main stream transect the strips of water-bearing rock, then gauging sections should be installed at the upstream and downstream boundaries of the permeable zones. An area of this type is usually more difficult to analyse than an area in which the draining streams and permeable zones coincide.

In areas that have been subjected to intense folding and faulting, steeply dipping impermeable rocks commonly abut permeable rocks (Fig. 7.1.1.1a, II.2). In this case, the ground-water flow boundaries may be the flanks of the relatively tight folds and drainage is to streams that occupy the troughs.

Observation wells are needed to measure seasonal changes of ground-water storage in the weathered overburden in all basin types discussed above. These wells should ideally be located along lines that cross the river valleys.

Values of specific yield and transmissivity that are obtained at a particular site (by pumping tests, finite difference methods, etc.) for typical rocks of mountain-folded areas may be of limited use when applied to other parts of the area or basin. The predominant fracture or fissure control on both water movement and storage in these rocks coupled with the commonly random orientation of fractures may make values for these parameters useless for areal considerations or calculations unless values from a large number of determinations can be averaged.

In certain basins, an important factor may be the relative values of specific yield and transmissivity of a tectonically disturbed zone as compared to an undisturbed zone. This comparison should be made only by use of adjusted values considered to be representative areally for each zone.

In outlets to mountain valleys the value of underflow should be estimated from water-balance data. If the estimated underflow rate is less than about 15 per cent of the stream discharge, more precise calculations are not needed. It is important in areas of this type to install wells on the lower and upper river terraces, in intermontane valleys, and in zones of tectonic disturbance. The numerous springs typical of such areas serve as additional points where data useful to the understanding of the ground-water régime may be easily obtained.

7.1.1.3 **Permafrost areas**

Key points that may be selected, dependent on the geothermal conditions of perennially frozen rocks, include (Fig. 7.1.1.1a, III): the junction of the seasonal thawing or freezing layer with perennially frozen rocks (sect. A); the occurrence of thawing soils and rocks in a stream valley (sect. B); in an interfluve talik (sect. D); under conditions of discontinuous frozen ground (sect. E); in a talik under a lake (sect. F), etc. The characteristic features of the unconfined ground-water balance for these plots are given in the notes to Figure 7.1.1.1a.

The method of computing the water balance for these plots is as follows:

1. The unconfined ground-water balance is computed for individual subzones or plots in which the role of the separate elements of the water balance is different.
2. For the seasonal thawing layer, the sum of the values of ground-water storage changes and moisture storage changes in the zone of aeration ($\mu\Delta H + W_2$) is determined. This total is derived from the difference between the final and initial moisture reserves in the zone from the land surface to the minimal water table during the period of observation.
3. During the warm period of the year, the rate and variability of ground-water flow to streams (if present) are dependent upon the rate of thawing of the permafrost.

The following features typify permafrost areas: (a) close relationship between the occurrence, recharge, and movement of ground water and the special conditions of perennially frozen ground and seasonally thawing rocks and taliks; (b) occurrence of ground water in the solid phase over long periods of time and changes of the state of water; (c) accumulation of surface waters and discharging ground water in the form of surface ice bodies; (d) close relationship between the ground-water and temperature régimes of the active layer and the zone of deep circulation; (e) non-uniformity of the structure, heat and water régime of the permafrost zone; (f) variability of water transmissive and storage properties of saturated permafrost rocks over the period of a season or a number of seasons.

The foregoing features are all related to the thermal régime peculiar to the zone. Thus, the study of the water régime of permafrost rocks cannot be accomplished without a parallel study of the temperature régime of the water-bearing rocks and the heat balance on the land surface. The following guidelines are given for study of the unconfined ground-water régime and balance in permafrost zones:

1. In studying the water balance in the permafrost zone it is necessary to recognize that this zone is not continuous even though the perennially frozen strata may be very thick. The permafrost zone is discontinuous, due to multiple taliks commonly associated with tectonic disturbances. Valleys and the deep circulation of ground waters in the subzone of insular permafrost play the principal role in the processes of heat and water exchange.

   The anomalous rise in the temperature of rocks and ground waters, as well as the peculiarities of their chemical composition (e.g. transition from calcium bicarbonate to sodium-cloride waters), is characteristic of flow through taliks.

   The presence of intra-permafrost and infra-permafrost ground waters adds to the complex pattern of natural water circulation in the permafrost zone. The role of baseflow in this complex pattern can be assessed by observations of the temperature and chemical composition of river and spring waters continued through the year.
2. On the basis of a hydrologic reconnaissance of basin conditions, one or two (or more if required by basin complexity) key plots should be chosen adjacent to a major or tributary stream for subsequent water-balance and thermometric observations.

   The following types of observation wells and other instruments are installed on each key plot:

   (a) A line of five or six observation wells screened in non-frozen aquifers (e.g. in

intra- or infra-permafrost ground-water zones) if they occur at depths of no more than 50-100 m from the land surface and if they exert an influence on the streamflow régime. These observation wells should be spaced several hundreds of metres apart. Shallow boreholes or dug wells (diameters about 1 m) are installed near the deeper wells to monitor the shallow ground water which overlies the active layer of permafrost.
   (b) Thermometric groups of mercury thermometers or thermistors at depth intervals of 0.5 m down to a depth of 5 to 10 m near observation wells.
   (c) A gauging station for observation of the stage and discharge of the river.
   (d) A network of snow-survey poles for observations of the depth and density of snow cover in winter.
   (e) A meteorological station.
3. Observations should be made of the volume and density of ice bodies formed as a result of discharge of ground or surface waters. For this purpose a topographic survey is conducted. A geodetic network of poles is established on the ice bodies to aid in recording the dimensions of those bodies and the density of ice and snow is determined periodically to estimate the equivalent water volume.
4. Ground-water discharge from springs can be monitored during the warm period of the year by simple water diversion devices such as flumes, pipes and spillways.
5. During the observation of water levels and spring discharges, water and air temperatures should be determined and water samples taken from typical springs and wells at different seasons for chemical analysis. Analyses should also be made of river water, snow, and melted ice from typical ice bodies. If free gas is emitted its composition should be determined.
6. Another important determination is that of the moisture deficiency of thawing soils in the zone of water-level fluctuations during the warm period of the year. Soil samples are taken at intervals from the land surface to the water table and the natural moisture and maximum moisture storage capacity are determined. These indices usually vary during the year, so that the determinations should be made two or three times a year on key plots during the first few years of observation.
7. For detailed studies of the ground-water régime under permafrost conditions, it may be necessary to select on representative basins several additional observation wells located (a) on flood plains and above-flood-plain terraces some tens to hundreds of metres from the river, (b) on long gentle slopes with different exposures and depths of perennially frozen rock occurrence, and (c) on plateaux and mountain tablelands.
8. Observations of ground-water levels under permafrost conditions are complicated because the water column in wells is often frozen. To prevent the wells from freezing, non-freezing liquids such as kerosene or oils may be placed in them. Levels measured under such conditions must then be adjusted by a correction factor that is dependent upon the difference in specific gravity of the non-freezing fluid and the water in the well. This method cannot be used where the non-freezing liquid can escape from the well and contaminate the aquifer. An alternate method of making measurements in wells that freeze is to use pressure transducers and thermistors that have been lowered below the frozen part of the well. However, this prevents observations of the chemical composition of ground water. The wells are sometimes warmed before recording the water level but a thorough pumping is necessary before sampling for chemical analysis.

7.1.1.4 Submontane terrains and valleys in arid regions

Climatically this zone is arid. The following subzones are distinguished within the zone (Fig. 7.1.1.1a, IV): (a) the pebbled (coarsest) part of the alluvial fan, (b) the middle part of the alluvial fan, (c) the subzone of unconfined ground-water discharge, (d) the subzone of secondary submergence and evaporation of ground water, and (e) the subzone of alluvial ground water.

Close relationship between the water balance of the subzones is typical of these regions. Some outflow elements of the water balances of one subzone are the inflow elements of other subzones.

The pattern of the unconfined ground-water balance varies for each subzone and the values of infiltration of precipitation, ground-water evaporation and recharge are very variable (see notes to Fig. 7.1.1.1a, IV).

The principal feature of this zone is a clearly defined microzonation in the geomorphology and lithology of submontane terrains and the sharp variation of hydrogeological conditions from the elevated areas to the lower plains.

Unconfined ground water in the pebbled part of the alluvial fan occurs typically at considerable depths (20 to 50 m and more) and the permeability of the coarse clastic rocks is very high. Ground water is recharged by seepage from natural streams and from artificial sources such as irrigation canals. Seepage losses from open channels may reach 50 per cent or more of their flow rates. Water-table evaporation is absent and water-table fluctuations depend on the non-uniform (in time) recharge from above. The amplitude of water-level fluctuations may be significant throughout the year, reaching 2 to 3 m. The main loss item in the water balance is ground-water outflow.

The principal operations recommended for carrying out studies in representative small basins located on alluvial fans are the following:

1. Two or three observation wells are installed on the line normal to a surface stream that produces clearly defined water-level fluctuations in order to calculate aquifer parameters (hydraulic diffusivity, specific yield, transmissivity). This direction must (as far as possible) coincide with the main direction of ground-water flow, if the latter is essentially one-dimensional.
2. The method of finite differences may be applied for estimation of unconfined ground-water recharge. In this case observation wells are drilled along the lines coinciding with the direction of ground-water flow. However, in view of the relatively great depths to water and the technical difficulties in drilling deep wells in gravel deposits, it may be advisable to use analytical solutions of differential equations for semi-finite and finite ground-water flows. Such an approach requires fewer observation wells.
3. Seepage losses from canals or other losing streams may be determined from discharge measurements in typical sections.
4. Observations of other water-balance elements are obtained at selected typical localities. The number of wells needed for this purposes is usually no more than 10 for a typical small basin.

   The evapotranspiration value less the condensation value ($Z - K_1$) is determined either by weighing lysimeters or by the heat-balance method. Changes in moisture storage in the zone of aeration are found to depths of about 2 to 3 m, below which moisture storage commonly changes little during the year.
5. Knowing the change in unconfined ground-water storage during the year, $(\mu \Delta H)_{av}$ and the average ground-water recharge from above $(W\Delta t_{av}$, for the balance area of a representative basin the value of the local ground-water recharge for the balance area can be determined:

$$\left(\frac{Q_2 - Q_1}{F}\,\Delta t\right)_{av} = (W\,\Delta t)_{av} - (\mu \Delta H)_{av} \qquad 7.1(8)$$

Middle part of alluvial fans. The middle part of alluvial fans is situated between the topographically high pebbled zone and the lower discharge zone of spring occurrence. Under natural conditions this is a flow-through zone receiving little additional recharge from seepage and losing little water to evaporation (as the water table is usually deeper than 4-5 m). Accordingly, single observation wells may be installed for determination of change of ground-water flow in this zone, both in space and time.

Subzone of unconfined ground-water discharge. The subzone of unconfined ground-water discharge is located on the periphery of the sloping part of the alluvial fan and results from the intersection of the water table with the land surface. Here it is common for ground-water discharge to result in numerous springs and considerable evaporation losses in cases of very shallow ground water. The other part of the ground-water flow (Fig. 7.1.1.1a, IV) submerges under clay loam alluvial sediments and becomes confined ground water.

The principal objective of water-balance investigations in this subzone is the determination of the portion of ground water that discharges here and the portion that becomes confined and flows on to the next downdip subzone, the submontane plain.

In addition to systematic observations of the discharge rates of typical springs in the subzone, the investigation should include mass hydrometric records of spring flow rates made two or three times a year. This work is undertaken at selected points within a system for which the width and the area of the recharge region situated upstream should be known.

If the discharge zone is large enough and considerable unconfined ground water loss by evaporation takes place, then observations of evaporation from the water table are necessary in addition to the hydrometric observations mentioned above. For determination of unconfined ground-water evaporation, lysimeters loaded with soil monoliths and saturated with water at different depths to water (such as 1, 2, 3 and 5 m) are used.

Subzone of secondary submergence and evaporation of ground water. The subzone of secondary submergence and evaporation of ground water is located on submontane plains that are composed predominantly of loamy alluvial sediments. They have intercalations of sand strata and are underlain by sand-gravel beds that contain confined (artesian) water that has moved downwards from the updip subzone. The water table occurs in loams and artesian water in the underlying sands. The artesian water that moves vertically upwards becomes unconfined ground water, some of which is discharged by evaporation.

For this circulation pattern, the vertical water exchange between the loams and underlying sands and the parameter K/μ (where K is the hydraulic conductivity of the upper layer in the vertical direction and μ is the specific yield of the rocks) should be determined. Wells are installed in unconfined ground water and in the underlying confined ground water, the number of observation points being dependent on the degree of homogeneity of the confining loam. These wells function as stage piezometers; data from them supplement the data from the normal observation-well network.

Subzone of alluvial water. The subzone of alluvial water is characterized by the occurrence of fresh ground water of vadose origin and of highly mineralized unconfined ground water that has formed as a result of intensive evaporation.

Fresh unconfined ground-water lenses usually occur under bare eolian ridges. In such cases, observation wells are arranged along lines or in groups of 4 or 5 to supply water-level-fluctuation data for the calculation of ground-water recharge and are located on elevated or lower parts of the relief and on slopes of different exposure.

In addition to observations of infiltration and evaporation (by use of evaporation pans, lysimeters, soil-moisture meters, neutron moisture meters, etc.), observations should be made, if possible, of water-vapour pressures in the near-surface layer of the atmosphere and in the zone of aeration.

7.1.1.5 Areas of alluvial and outwash plains

This zone is characterized by the shallow occurrence of unconfined ground water and frequent changes in time and space of the moisture deficiency and water yield of soils of the zone of aeration. It is necessary in compiling the unconfined ground-water balance to consider carefully the changes in moisture storage from the land surface to the water table (i.e. $\mu \Delta H + W_2$), and the vertical water exchange between layers. The downward flow of water from upper to lower layers is characteristic to elevated areas. The reverse is typical in lower areas, which accounts for the moistened land surface in such areas. The formation of ground-water flow that replenishes streams is more complex. The diagram of hydrograph separation is similar to the diagram for heterogeneous conditions of the platform type (Fig. 7.1.1.1a, I.2 and V).

These areas are characterized by an average annual excess of infiltration from precipitation over evaporation from the water table and land surface. Average recharge thus exceeds discharge and water accumulates on the land surface forming swamps over most of the year. The studies of the ground-water régime and balance in such areas are of primary importance to their reclamation, i.e. control by drainage of their excessive moisture.

In each element of relief or ground-water zonation (Fig. 7.1.1.1a, V) test drilling should be carried out to ascertain the distribution of ground waters of various types and the interrelationship between the ground waters and rivers, lakes, and swamps. Some of the test wells should be completed as observation wells in areas where there are significant vertical differences in aquifer heads.

After test drilling, key plots are chosen in areas that are typical of the basin and in those subject to future reclamation and a network of observation wells installed. Two adjacent plots with common conditions of infiltration and evaporation of ground water over the area but with different drainage conditions are selected for the determination of hydrogeological parameters.

For determination of changes in ground-water storage and recharge from water-level fluctuations it is necessary to ascertain the values of moisture deficiency and specific yield for soils in the zone of water-level fluctuations. In determining the average changes in ground-water head and storage over a basin it is important to make preliminary comparisons of these values to the depths to water in various geomorphological elements of the basin. Thus a relatively dense network of observation wells in various relief forms must be measured during the first 2-3 years of observation to allow this comparison. After correlations have been established the number of observation wells may be reduced.

In drainage reclamation areas, it is important to monitor the functioning of the drainage network. Such monitoring should include the following: (a) determination of discharges of drains, collectors and channels; (b) determination of the changes in the unconfined ground-water régime (water-level position, chemical composition of water and variation of régime elements) caused by drains of various types; observation well lines are arranged normal to drains of various types and the well lines are extended to the limit of the zone of canal influence; the distance between the wells will be governed by local conditions but generally is several tens of metres or less; (c) for detailed investigations of the efficiency of drainage systems, stage piezometers should be installed and screened at various depths; water-level data from such systems allow construction of ground-water flow nets near and between drains.

7.1.1.6 **Karst areas**

Special attention should be given in karst areas to the circulation and flow of ground water in vertical zones in addition to the usual régime observations.

According to Sokolov (1962) the typical zones of karst ground-water occurrence are (Fig. 7.1.1.1a, VI): (a) the zone of aeration characterized by downward infiltration of water of atmospheric origin, sometimes perched water; (b) the zone of seasonal water-table fluctuation characterized by horizontal movement when levels are high; (c) the saturated zone below the water table, this zone usually receiving recharge locally and discharging to local streams; (d) the zone of deeper circulation where water moves from recharge areas that may be outside the local basin to distant areas of discharge outside the basin.

In addition, 'syphon circulation' along isolated channels in zone (c) is possible. This is a special case of confined ground-water flow in karst.

Surface waters may play a highly variable role in water-balance determinations in karst areas as certain reaches of a stream may be gaining ground water and others losing water. These conditions in a reach may change in time as conditions change in the ground-water régime. This affects water-balance determinations because a change in conditions changes the algebraic sign of the particular element in the balance equation.

For relatively large drainage basins (Fig. 7.1.1.1a, VI) with relatively steady flow conditions between sections 1 and 2 the base flow is determined from the formula:

$$Y_b = \sum_T W_{\mathrm{w.av.}}\, \Delta t + Q_{\mathrm{asc}} - \sum_T (\mu \Delta H)_{\mathrm{w.av.}} \qquad 7.1(9)$$

According to this equation the base-flow, Y_b, is the algebraic sum of the averaged value of recharge by infiltration of precipitation $\Sigma W_{\mathrm{w.av.}}\Delta t$, the upward flow of ascending water Q_{asc}, and the change of water in storage, $\sum_T (\mu \Delta H)_{\mathrm{w.av.}}$, in the basin.

The unconfined ground-water balance and the total water balance for karst areas are determined from the equation:

$$\sum_T (\mu \Delta H)_{\mathrm{av.}} = X - Z + K_1 - W_1 + Y_{1,\mathrm{ov}} - Y_{2,\mathrm{ov}} - Y_b + Q_{\mathrm{asc}} \qquad 7.1(10)$$

where $Y_{1,\mathrm{ov}}$ is the overland flow into the basin.

The determination of discharge of ascending waters, Q_{asc}, which may move through the siphon system of karst channels may be difficult. The relationship between zones (c) and (d) is also complex. There is some exchange of water between them but the latter serves mainly as a zone of transit for deep circulation of karst waters.

The examples considered and typical diagrams presented for certain representative basins cannot cover the wide range of naturally occurring conditions that affect water balance formation. Before detailed investigations are made of an individual basin of any type, construction of a water balance should be attempted from observations of natural conditions. In the course of detailed studies, the construction is improved and the balance becomes more precise.

A good deal of attention is given to surface-water observations in karst areas because of the close interrelations and frequent interchanges between surface and ground waters. When beginning an investigation in a karst basin, temporary gauges and measuring sections should be set up on streams to reveal the basin areas where karst aquifers are recharged by river waters (losing streams) and rivers drain the aquifers (gaining streams). After identifying basin plots which are typical as regards the interrelationship of ground and surface waters and establishing the general pattern of water movement in the main channel and tributaries, the sites where surface-water observation stations should be installed to obtain long-term records are chosen. The number of gauges and measuring sections needed depends on the complexity of the region.

To determine hydrogeological parameters of karstic rocks, lines of parametric observation points are arranged perpendicular to the main channel of the river, extending preferably to the ground-water divide. In addition to wells, the observation points may include sink holes and other karst forms of vertical cavities that penetrate to the karst water table and allow observations of the ground-water régime to be made.

Vertical zonation of karst water is studied by installation of groups of stage piezometers at various typical environmental settings in the basin. These settings would include areas near stream channels, on topographic highs within the basin and near the basin divide.

To determine the change of karst water storage it is useful to establish single observation points at different topographic positions in the basin.

Observations should be made of the flow rates, temperature and chemical composition of water from the larger springs in a basin. These observations are begun during reconnaissance of the area and continued for selected springs. Typical springs should be observed systematically for a number of years.

Tracer studies may be used in defining ground-water flow patterns in karstic rocks, particularly to identify the interrelations between ground and surface waters in discrete areas underlain by highly permeable rocks. As tracers, colouring agents such as fluoroscein, rhodamine or other dyes can be used and common table salt (NaCl) may be an effective tracer in some fresh-water environments.

7.1.2 *General provisions for location of observation wells*

Certain requirements as to the location of observation wells should be observed when using methods of hydrodynamic analysis for calculating aquifer parameters and when compiling the ground-water balance for the unconfined ground-water régime.

Consider first the elements of an observation well network installed for use in the ground-water investigation of a small experimental basin. Figure 7.1.2a shows a network in such a basin with an area of about 100 km^2 underlain by fluvioglacial deposits. The network of observation wells has the following elements: (a) parametric observation wells installed along short lines normal to rivers, canals, reservoirs and other water bodies to monitor ground-water level changes in relation to surface-water level changes; (b) calculation lines of observation wells (for determining the infiltration of precipitation and evapotranspiration of ground waters) positioned along the main direction of ground-water flow; (c) groups of calculation observation wells to allow determination of the same water-balance parameters as (b) but under conditions of two-dimensional flow; (d) single observation wells installed in various areas to supply data for construction of ground-water contour maps and calculation of changes in ground-water storage; (e) groups of several wells screened to monitor various points in the flow system and various aquifers at different depths when water-bearing beds are heterogeneous, these wells being installed to determine the interrelations between aquifers; (f) water-balance plots and sites with a dense network of observation wells, test instruments and units for experimental observation of water-balance elements at the land surface, in the zone of aeration, and in the zone of ground-water flow.

Parametric observation wells should be located near the boundaries of ground-water flow (rivers, canals, reservoirs) so that water-level fluctuations in them are not less than 0.3-0.5 of the amplitude of the water-level change at the flow boundary. The distance from such a boundary to the most distant parametric well should be about 0.5 $\sqrt{at}$ to 1.0 $\sqrt{at}$, where a is the hydraulic diffusivity (T/S) and t is the time over which the most significant change in water level at the flow boundary occurs.

On the basis of the results of a reconnaissance ground-water survey preceding any water-balance studies, the basin territory is zoned according to the major factors that influence the ground-water régime. In such zoning meso-and microrelief, vegetation, soils, lithology, degree of natural drainage of aquifers and other local factors are taken into account. Typical sites are then chosen where calculation lines or groups of observation wells are installed.

Each calculation line or group of observation wells should afford data on several hydrologic parameters needed for water-balance determinations. In the case of one

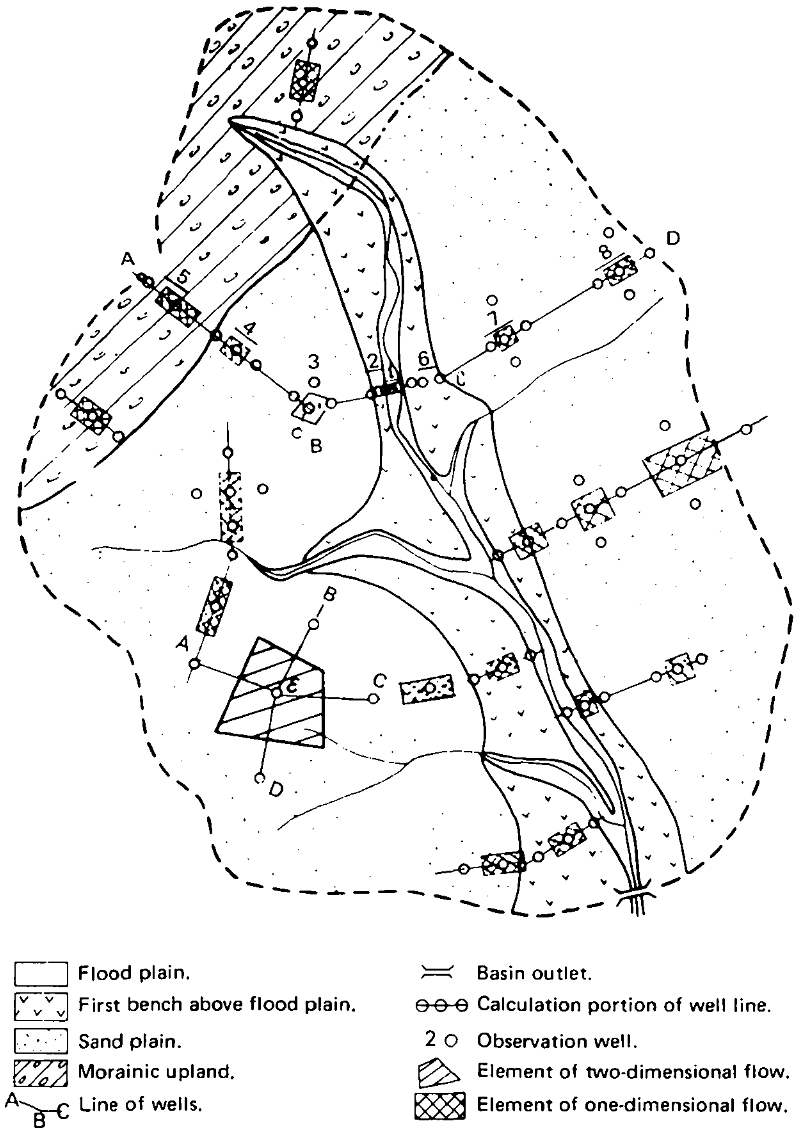


FIG. 7.1.2a. Location of observation wells in an experimental river basin.

dimensional flow—that is, when parameters are being measured along a flow line—no more than three wells located along the flow line are needed to constitute the calculation line of observation points.

To monitor two-dimensional movement, it is necessary to add an additional well on each side of the calculation line to form a calculation group of five observation wells.

Such important hydrologic parameters as infiltration of precipitation and evapotranspiration of ground water will vary as the thickness of the zone of aeration varies, assuming that all other conditions remain constant. However, if these parameters have been determined from data from calculation lines or groups of wells, the computed values may be extrapolated over most of the area if sufficiently detailed water-level contour maps are available. Most of the data used in the preparation of such maps are obtained from single observation wells constructed to supply water-level data.

Water-level fluctuation data from observation wells nearest the basin divide or the draining stream should be used to calculate hydraulic diffusivity, T/S. When calculation wells are installed by a stream that does not completely penetrate the aquifer, the first such well should be installed at a distance of about 1.5 times the thickness of the aquifer from the stream.

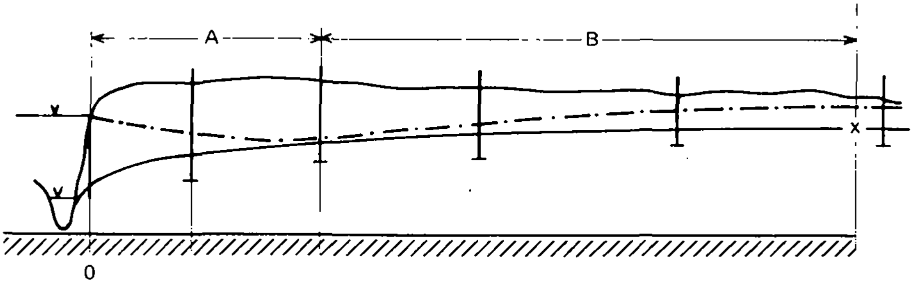
At calculation sites near streams, an observation well at the water's edge and a staff gauge in the stream should be installed to supply data on the degree of hydraulic connexion between the water-table aquifer and the stream.

It may be necessary to install groups of observation wells screened in various water-bearing zones when a highly heterogeneous aquifer is encountered in a basin. Data from such wells permit interpretation of aquifer interrelationships from a comparison of water levels in the various zones.

The preceding paragraphs have discussed the types of observation wells that should be installed in an experimental basin to supply data for calculating aquifer parameters and compiling ground-water balances. The following paragraphs include more generalized comments on the location and arrangement of observation wells in various environments.

In areas that contain no draining streams or concentrated sources of recharge, observation wells are located at various distances from each other and in various patterns depending on the location of surface factors (vegetation, microrelief, etc.) whose influence is under investigation. Conversely, if the area of study is of the more common type found in humid regions—that is, a basin drained by a perennial stream—the line of observation wells should extend beyond the influence of the flow boundary to the ground-water divide. This installation scheme is shown in Figure 7.1.2b.

If a line of observation wells interesects forest tracts, the wells should be located on forest edges, in the forest and in clearings in the forest. Calculation sites, each having 2-3 wells, make it possible to compute a ground-water balance which considers the changing moisture conditions at the land surface. In this case, the hydrologic effects of forestation can be appraised more reliably.



A. Zone of bank type of unconfined ground-water régime.
B. Zone of divide type of unconfined ground-water régime.
——Profile curve for the minimum water-table position.
–·–Profile curve for the maximum water-table position.
⊥ Observation well.
x Ground-water divide.

FIG. 7.1.2b. Diagram of hydrodynamic zones.

If ground-water occurrence is studied in an area bounded on both sides by rivers, canals or reservoirs, observation wells should be located not only near these boundaries of the flow, but also in the interfluve.

In Figures 7.1.2c and 7.1.2d, diagrams of observation-well locations in an interfluve and on a piedmont plain are given. Groups of three observation wells are shown in these figures to which the finite-difference method of computing balance elements may be applied. Additional observation wells for investigating underlying interstratal artesian waters are shown in Figure 7.1.2d.

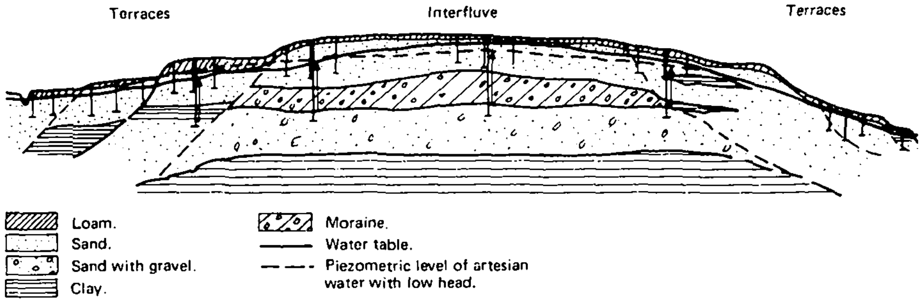
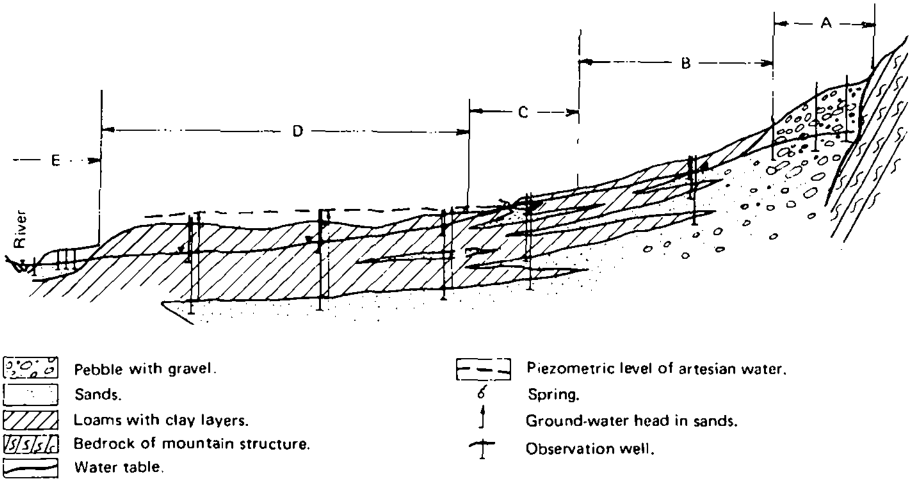


FIG. 7.1.2c. Schematic profile along an interfluvial line of observation wells on a balance plot.



Legend
A. Pebbled (coarsest) part of alluvial fan—area of recharge by infiltration to ground water.
B. Middle part of alluvial fan—transit zone of ground water.
C. Periphery of the upper alluvial fan and zone of ground-water seepage.
D. Zone of secondary sinking of ground water and vertical water exchange between loams and underlying sands.
E. River valley:

FIG. 7.1.2d. Diagram of inclined piedmont plain.

7.2 Installing and testing observation stations

When studying ground-water conditions wells and springs serve as observation points. It is usually undesirable to use large pits or production wells for water-table observations inasmuch as their large diameter and production effects may distort the fluctuations in, or the shape of, the ground-water body, resulting in erroneous interpretations of the areal conditions.

Many publications treat in detail the installation and equipping of observation and test wells.

7.2.1 *Wells*

Wells installed for the purpose of studying ground-water conditions in an area are termed observation wells or piezometers.

Observation wells, like test wells, can be drilled by any one of several different methods using drill rigs of various designs, the most common being the rotary and the cable-tool percussion types. Regardless of the drilling method, care should be exercised to minimize impairment of the natural permeability of the water-bearing materials penetrated, especially when mud fluids are used during the drilling operations. Considerable attention must then be given to well development after drilling, to remove any fines brought into the water-bearing zones during the drilling process.

Well drilling in unconsolidated deposits is usually accompanied by casing installation. Apart from strengthening the walls of the well the casing serves also to isolate individual aquifers from each other. In arid regions where the dissolved-solids content of ground water may change considerably with depth, the installation of well casing makes it possible to carry out layer-by-layer testing of the individual aquifers as the drilling progresses. In consolidated rock formations wells may be drilled and completed without casing.

Until recently metal pipe has been the primary material used for well casing. With the advent of polyethylene plastic pipe, however, this material has found wide application where formerly asbestos-cement pipe would have been used. Non-corrosive, non-metallic pipe such as fibre glass is appropriate for casing observation wells in zones which yield waters that are especially corrosive.

For water-level observations of shallow water-table aquifers in loose clayey sand deposits wells may be constructed using drive-point screens (Fig. 7.2.1a), mounted on a length of pipe. The screen proper may be as much as 1 m long and commonly consists of a perforated pipe wrapped with a suitable mesh screen. The sharpened drive point is somewhat larger in diameter than the screen section. Drive-point screens are usually not installed at depths greater than about 10 m.

Observation wells are commonly of small diameter, primarily because the drilling of a small-diameter well is cheaper. However, the well diameter must be large enough to permit a measurement of the water level without difficulty, to install a recorder float, to use a water sampler if changes in ground-water chemical composition are to be studied and to allow the periodic removal of accumulated sediment and trash. Considering these factors wells are commonly constructed with inner diameters ranging from 5 to 15 cm.

Proper design of an observation well must ensure reliable isolation of the aquifer to be observed from any other aquifers present, as well as from precipitation which might seep downward outside the well casing. The design must also ensure good contact with the aquifer to be studied. The following criteria are worthy of note in promoting good observation-well design (Fig. 7.2.1b): (a) screen that ensures free interchange of water

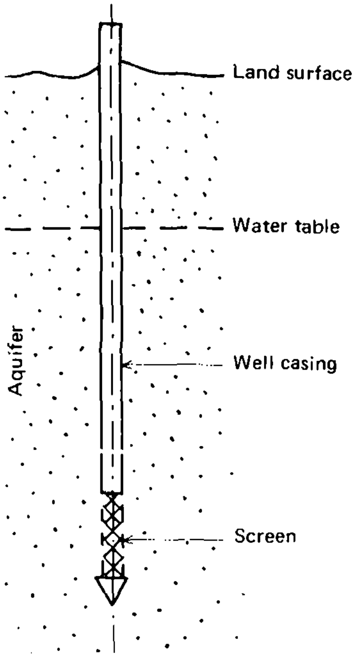
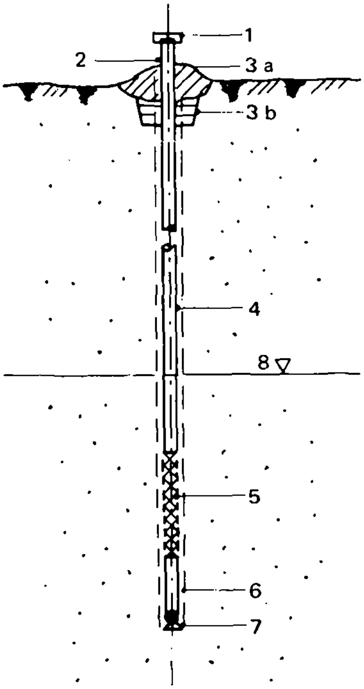


FIG. 7.2.1a. Drive point screen and casing for observation well.



1. Casing cap.
2. Pipe nipple extension of casing.
3. Combined cement/clay seal (a, cement cushion; b, clay seal).
4. Casing string.
5. Well screen.
6. Debris sump.
7. Plug seal.
8. Water table.

FIG. 7.2.1b. Design features of observation well.

between the investigated aquifer and the well; (b) casing to which the screen may be fitted by any one of several different methods; (c) casing extension above ground, preferably using a pipe nipple made of material having low thermal conductivity, to a height of 0.7-1.2 m; (d) casing cap fitted with a locking device to protect the well from accidental or premeditated obstruction and to prevent precipitation from entering the well directly; (e) the measuring point from which measurements of water level are made should be related by levelling to a near-by fixed reference point.

7.2.1.1 **Well finishing and sealing operations**

To prevent downward percolation of precipitation or surface run-off along the outside of the well casing a seal is made at the land surface (Fig. 7.2.1b). The seal can be of clay, cement grout or combined material. To construct a good seal a dug pit is recommended with a cross-sectional area of about 1 m^2 and about 1 m deep. This is backfilled with solid bentonite and firmly compacted with cement grout or with clay in the lower part, and grout in the upper part (Fig. 7.2.1b). The latter-type seal is obviously stronger and more durable than one made only of clay. At the land surface the backfilling should be mounded up to form a small cone-like elevation around the extended casing to provide good surface drainage away from the well.

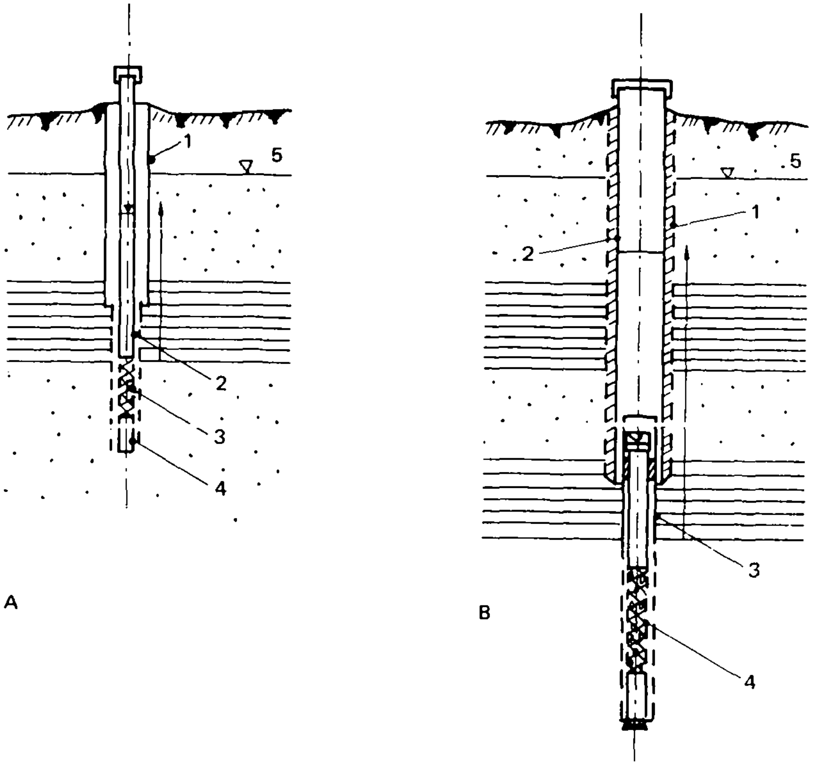
Often it may be desired to develop a well in an aquifer below the first water encountered. In this event it is necessary to case out and seal off the upper aquifer(s) so that the chemical quality and piezometric head in the aquifer under study are not impaired. The seal can usually be accomplished by one of the following methods:

1. If the aquifer or zone under study is overlain by a clay layer not less than 2 to 3 m thick, sealing can usually be accomplished by simply driving or pushing the outer casing shoe into the clay until it is solidly seated (Fig. 7.2.1.1*A*) Drilling can then proceed into the desired aquifer. When no natural clay layer is present a temporary seal may be constructed by filling the bottom of the well casing and drill hole with viscous clay and using a cone-shaped plug to press the clay out into the annular space surrounding the casing.
2. The most reliable method for sealing off desired zones is through the use of cement grout (Fig. 7.2.1.1*B*).

An inexpensive and simple grouting procedure involves: (a) running the casing into the drill hole, mixing enough grout to fill the lower 12 to 15 m of the hole and placing the grout in the bottom of the casing with a dump-bailer or by pumping; (b) with the grout in place, lifting the string of casing 12 to 15 m off the bottom of the drill hole, depending on the amount of grout placed, keeping the lower end of the casing submerged in the grout; (c) capping the casing, after filling it with water, and lowering it to the bottom of the drill hole thereby forcing most of the grout up the annular space around the outside of the casing. The cap should not be removed from the casing until the grout has set. If in step (b) it is anticipated that there may be difficulty in lifting the casing the water filling can be done without lifting. The casing should then be capped with provision for a pump connexion so that additional water can be forced in under pressure. Only a measured water volume equal to the quantity of grout to be displaced should be pumped.

The success of the sealing and grouting procedure should be checked by bailing the water out of the casing and observing whether or not the casing refills. After the grout has set for about 24 hours drilling can proceed to the next or desired aquifer.

If the hydrologic conditions are to be investigated in a vertical sequence of several aquifers, an observation well should be installed in each aquifer. However, it is possible to drill only one well to study the several aquifers if a well screen is placed in each aquifer and isolated from the others by grouting or other sealing methods. It should be noted that, for any multiple-screen installation, reliability of the procedure for sealing off the individual aquifers should meet the highest standards.



A. Sealing of casing in clay layer
1. Outer string of well casing.
2. Inner string of casing and screen.
3. Well screen.
4. Sediment sump.
5. Water table.

B. Sealing of casing by grouting
1. Grout-filled annular space outside casing.
2. Outer string of well casing.
3. Inner string of well casing.
4. Well screen.
5. Water table.

FIG. 7.2.1.1. Details of well-sealing operations.

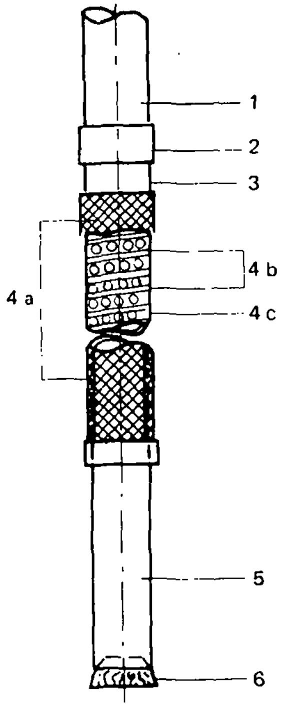
7.2.1.2 Screen design

Well screens should ensure free interchange of water between the aquifer and the well and should protect the well from sloughing and caving.

The design of screens has been discussed in the literature, as for example Bennison (1947), Gavrilko and Abramov (1956) and Gavrilko (1961).

A serviceable screen (Fig. 7.2.1.2a) consists of: (a) the screen proper which may be simply a perforated length of pipe with or without an open mesh or net jacket; (b) a sediment sump made of a blank section of pipe 1 to 2 m long secured to the lower end of the screen and into which rock particles entering the well through the screen may settle by gravity; and (c) a section of blank pipe above the screen. The sump is sealed at the bottom commonly with a metal or wooden plug or a back-pressure valve. The screen design is dependent upon the grain-size distribution of the porous rock materials and the chemical composition of the water.

Screens without a mesh or net jacket (Fig. 7.2.1.2b) may be installed in fissured consolidated rocks or in gravel deposits. Such screens are commonly made of metal or



Legend
1. Inner well casing.
2. Coupling.
3. Blank pipe above screen.
4. Screen proper: (a) open mesh or net jacket; (b) wire winding; (c) perforated pipe.
5. Sediment sump.
6. Bottom plug of metal or wood.

FIG. 7.2.1.2a. Well screen design features.

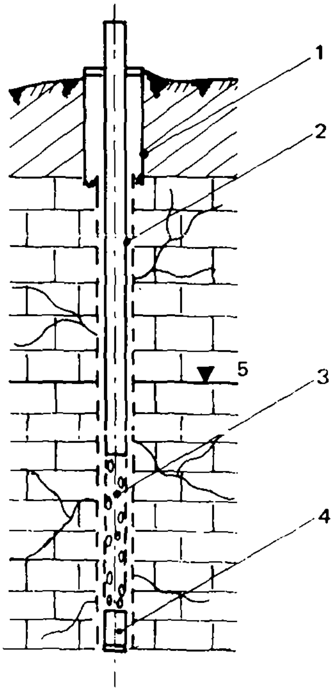
polyethylene perforated pipe. The perforations may be round or slotted types. Round openings are commonly arranged in draughts-board fashion at the apexes of equilateral triangles, the base of which is called the pitch of the perforations. Depending on the diameter of the openings and the desired screen capacity the pitch may range from 10 to 20 mm.

Screens with mesh or net jackets are one type used in sand deposits (Fig. 7.2.1.2a) and consist of a structural core, a wire winding and an outer mesh or net. The core is commonly made of perforated pipe. Wire 2 to 3 mm in diameter is wound spirally around the core with a pitch in winding of about 30 to 60 mm.

The wire wound on the core ensures that the outer mesh jacket will be held out away from the core and therefore will be less subject to clogging; water will move through more of the mesh open area and the screen will function more efficiently.

The choice of mesh, with respect to type, size and arrangement of openings, depends on the texture of the water-bearing sand. For uniformly graded sands a mesh with openings capable of admitting 40 to 60 per cent of the sand particles is commonly chosen.

A type of screen which offers the advantage of maximum open area per unit length is the continuous-slot-type wire-wrapped screen. This screen is formed of wire of a number of different cross-sectional configurations wrapped spirally around a series of equally spaced rods which extend the length of the screen. The wrapping is either a press fit into slots in the longitudinal rods or welded to them. Since there is no perforated pipe base the area of the slots becomes the effective area through which water can pass into the well.



Legend
1. Outer well casing.
2. Inner string of casing and screen.
3. Screen section of perforated pipe.
4. Sediment sump.
5. Water table.

FIG. 7.2.1.2b. Screen for use in consolidated rocks.

Gravel packing may be necessary if any of the foregoing screen types is to be placed in a fine or silty sand or a poorly sorted sand and gravel. The gravel pack is simply an annular layer of artificially graded coarser material placed to surround the screen to a minimum thickness of from 35 to 70 mm.

Basket screens may be fabricated in lieu of gravel packing to accomplish a graduated transition from the fine-grained sand in the water-bearing formation to the desired larger openings in the screen core. The fabrication can be done in a variety of ways making use of readily available materials and special designs. A simple basket screen can be made by securing slender wooden laths (with cross-sectional dimensions of, say, 1 × 1.5 cm), at a parallel spacing of 50 mm, to a length of perforated pipe using a wire winding. An appropriate mesh jacket constitutes the outer covering and the annular space between it and the pipe core is packed with gravel and coarse sand graded in the desired manner. This screen has the advantage over a conventional screen with gravel pack in that it does not require a large-diameter borehole.

A disadvantage of screens with mesh jackets is the added complications in their manufacture, and their comparatively low open area for admitting water. The latter circumstance, however, is not of critical significance for observation wells inasmuch as the volume of water moving into and out of the well due to ground-water fluctuations commonly is not great.

All screens are subject to some extent to plugging through silting by mechanical or accidental processes, which thereby lowers their water-transmitting capacity. Screen-silting processes appear less significant with polyethylene and asbestos-cement pipes which may influence their choice over metallic pipe.

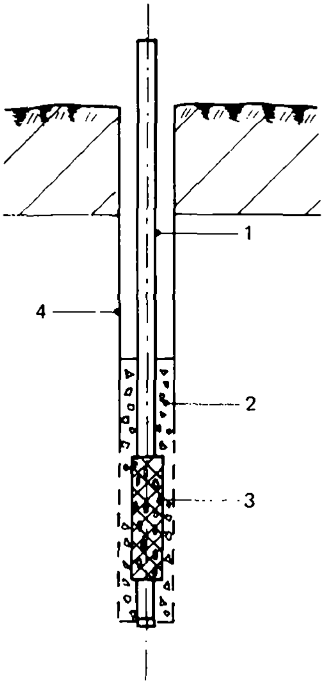
Anti-corrosive screens are manufactured from perforated polyethylene pipe covered with a plastic mesh jacket of fibre-glass, alkali-treated synthetic material or anti-corrosive metals. Blocks of porous ceramic can be fitted over perforated pipe, but one should bear in mind that iron or clayey suspensions in the water can lead to substantial clogging of the ceramic blocks.

7.2.1.3 **Screen installation**

The screen for an observation well is commonly installed at a depth that will ensure its remaining below the lowest anticipated water level. Exceptions to this may be for special investigations of the changes with time, layer-by-layer down through the saturated zone, of such ground-water-quality parameters as, say, the total dissolved solids.

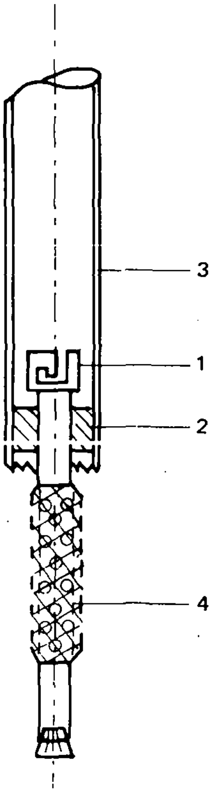
In this event a group of wells may be used with individual short-length screens (up to 0.5 m) placed at depths ranging downward from the highest known water table. Generally the screen length for observation wells need not exceed 2 m—a length that ensures good response of the well to water-table fluctuations. More precise specifications on screen length must be tailored to the local situation and to the purposes for which the water-level data are to be collected. If the aquifer is not of great thickness it may be appropriate to make the screen as much as 3 to 5 m long, installing it in the upper part of the aquifer but below the lowest anticipated ground-water level.

Procedures for installing well screens vary with the design of the well and with the method employed in drilling. The methods most commonly employed are briefly described in this section. Additional details are given in the previously cited references.



1. Continuous inner string of pipe with screen.
2. Gravel pack.
3. Screen.
4. Outer casing.

FIG. 7.2.1.3a. Installation of gravel-packed screen.



Legend
1. Coupling with quick-disconnect feature.
2. Seal.
3. Outer casing.
4. Screen.

FIG. 7.2.1.3b. Flush-fitted screen installation.

A procedure known as the pull-back method is commonly used where the well is drilled by the cable-tool percussion technique. The essentials of the method include lowering the screen, fitted with a lead packer ring at the top, inside the well casing to the full depth of the hole and then pulling back or lifting the casing a sufficient amount to expose the screen in the water-bearing sand. After the screen has been seated on the bottom a swedge block is run down the hole and repeatedly dropped on the lead packer ring to seal off the annular space between the bottom of the well casing and the top of the screen.

When installing a sand- or gravel-packed screen, the diameter of screen column generally must be from 80 to 100 mm less than that of the smallest diameter string of casing used in drilling the well. For proper positioning the screen column is provided with centring spiders or guide laths. After the screen-column installation the annular space between this column and the outer well casing is filled with graded sand or gravel. The outer casing is then pulled back or jacked up until the total length of screen is exposed (Fig. 7.2.1.3a). Commonly, the inner string of pipe carrying the screen remains in the well. However, to economize in situations where the screen is being set at considerable depth the inner pipe can be recovered and the screen left in a 'flush-fitted' (Fig. 7.2.1.3b) fashion. Two common means of accomplishing this are: (a) by equipping the pipe coupling just above the screen with two L-shaped notches (or similar quick-disconnect feature) that engage two machine claws on the inner pipe string or drill stem, or (b) by equipping this pipe coupling with a left-hand thread. In the first method, after the screen is lowered to the required depth and properly exposed and sealed the quick-disconnect feature is operated by turning the drill stem or pipe in the appropriate direction. In the second method, after the screen is properly set the drill stem or inner string of pipe is simply turned to the right until it separates from the coupling above the screen.

After installation the observation wells are developed by means of a bailer or a pump until the water clears. If the hole was made by the hydraulic-rotary method—using drilling mud—the cleaning and development process must ensure that as much of the mud and fine particles as possible are removed from the invaded zones. As a final step the well should be tested to make certain that it will be responsive to water-level changes in the aquifer.

7.2.1.4 Testing of observation wells

Testing is performed to reveal possible discrepancies between water-level fluctuations in an observation well and in the water-bearing formation that it taps. The testing should be undertaken immediately after the well is completed and repeated periodically.

In a variety of ground-water studies, including those leading to water-balance determinations, partially penetrating, cased observation wells finished simply as open holes are often used. Testing of such wells is conveniently done by pouring in a measured slug of water into the well and collecting data on the manner in which the water level returns to its original position.

To analyse the residual-head data following injection of a slug of water a semilog graph is constructed with values of $\log_{10}(H_0/H)$ plotted against values of time, t. The term H represents the residual or excess head in the well over the original static head at time t, where $t = 0$ at the moment the water is poured in, and H_0 is the initial value of H. If v is the volume of the slug of water and w_c is the cross-section of the observation well in the zone of water-level fluctuation, then $H_0 = v/w_c$. Experimental points on the semilog graph should lie on a straight line from which any co-ordinate values of $\log_{10}(H_0/H)$ and t may be taken. The log coefficient, α_c, for the observation well is then computed from the relation:

$$\alpha_c = 0.435 \frac{t}{\log_{10}(H_0/H)} \qquad 7.2\,(1)$$

Testing of a closed partially penetrating artesian observation well is accomplished by suddenly opening the well—allowing it to flow—and measuring the flow rate, Q_{nc}. If s_{nc} is the drawdown from the shut-in head to the level represented by the discharge outlet, then the log coefficient, α_{nc} is determined from the relation:

$$\alpha_{nc} = \frac{w_c\, s_{nc}}{Q_{nc}} \qquad 7.2\,(2)$$

An alternative method of testing an observation well involves pumping water from it and then observing the manner in which the water level recovers after the pumping is stopped, using Equation 7.2 (1).

7.3 Requirements of instruments and equipment for ground-water observations

Instruments and equipment used in observing the ground-water régime should be: simple and easy to transport and use; reliable and durable under widely varying climatic conditions; sufficiently accurate to obtain the information required; inexpensive; standardized, so that parts and units are interchangeable and international units of measurement obtained.

Observations of the ground-water régime are made in the field at both attended and unattended stations, and may be either periodic (for instance, daily, or once in 3, 5, 10, or 30 days or at any other selected interval) or continuous. Field instruments can be classified into two principal groups: portable or fixed. The latter may be further classified into attended and unattended, as well as semi-automatic and automatic with or without remote control. Automatic recording instruments have been developed which record the parameters being observed on perforated tape, so that the data can be processed by electronic computer.

Portable instruments for measuring the position of the water table must be easily carried, simple to use and accurate. For example, to achieve an accuracy in water-level observations of ± 10 mm for water depths of 100 m or more steel tapes are quite acceptable. These can be coated with a substance, such as carpenter's chalk, which will clearly mark the point to which the tape is wetted. The tape is lowered into the well to a depth greater than the depth to water, and the water level is calculated by subtracting the wetted length from the depth shown at the surface measuring point. A steel tape to which is attached a weight with a cavity in the base may also be used. When the weight strikes the water it makes an easily detected sound and the depth to water can then be read from the tape. In measuring greater depths to water, with relatively small amplitude of fluctuations under natural conditions, steel measuring cables may be used and may be left hanging in the individual wells. Corrections for deviation of the well from vertical, cable stretch and temperature may be necessary when measuring water levels of more than 100 m.

If the water table is at very shallow depth (less than 3 m) it may be convenient to use a portable staff with an electric contact which closes when it touches the water surface and activates a small light bulb fed from a dry-cell flashlight battery. Such a staff, provided with a micrometer adjusting screw, decreases the error of water-level measurement to about 2 mm.

A similar system, consisting of an insulated two-conductor electrical cable with a shielded electrode on the lower end and a milliammeter and battery across the terminals at the upper end, may be used to measure deeper water levels. Such a system is commercially available in the United States and United Kingdom in units which can measure water levels in excess of 150 m.

Instruments used for recording ground-water levels at continuous record sites should be easily serviceable and must be sufficiently precise to obtain the desired data. When making observations to show the influence of various meteorological factors upon ground-water levels, it may also be desirable to obtain daily graphs of atmospheric pressure, precipitation and air temperature to correlate with the water-level records. The recording-scale factor should be the largest possible consistent with an easily interpreted record. Automatic recorders with a weekly or monthly clock mechanism are commonly used; however, recorders are now available which allow the collection of records for several months. Multi-channel recording instruments are available that can register on single charts not only the variations in water level but other parameters as well.

Remote-recording gauges and water-level recorders play an important role in instrumenting water-balance test sites, in accessible locations, or observation points at distant locations. At test sites (plots) where there is a demand for systematic and frequent information on ground-water levels, a land-line communication system featuring a central control system may be a logical solution. It is essential that such an installation be standardized so that the numerous electric gauges and recording equipment will be compatible with the remote equipment.

From operational experience in a number of countries the utility of the wire-communication system for connecting remote stations is unquestioned as long as it is economically practical to run land lines to the observation points. For observation points more remote from the control centre, it is more practical to use telemetering equipment with transceivers operating in the high-frequency or short-wave band.

In the quest for a unified, standard or optimum plan of remotely controlled observations of ground-water levels, the following factors should be considered: (a) remotely controlled gauges should be restricted to observation sites in virtually inaccessible or distant locations; (b) radio transmission of information is important when changes in ground-water levels or chemical composition are significant, or when these changes can bring about disastrous effects requiring emergency control measures; (c) technical field inspection of the operating equipment cannot conveniently be made at monthly intervals; (d) the central observation unit should be capable of controlling up to 100 observation points including interrogating, reading and recording the data; (e) recording of the transmitted data should be automated, using graphical and digital form. The readings at selected intervals in the record should also be entered on punched or magnetic tapes at the remote site in order to prevent loss of record in the event of transmission problems.

The installation of remotely controlled devices for observing a ground-water régime should only be undertaken after a full evaluation of the technical and economic aspects of their application. If a river basin is not large and is visited frequently by observers and hydro meteorologists, it will generally be simpler and less expensive to use conventional equipment which is easier to handle and operate. For small seasonal changes in ground-water level, portable instruments may be the most effective means of making the observations, provided that correlation with meteorological conditions is not required. Furthermore, field visits always afford the opportunity for recording related observations, such as growth stages of vegetation, agricultural work, new industrial or irrigation development and land levelling or clearing that might affect the ground-water régime.

The water head in flowing wells can be measured using manometers or pressure gauges mounted on special well caps fitted with valves for temporarily closing the well.

Water levels in wells subject to freezing can be measured by means of electrical transducers of the pressure cell or membrane type, or a protective layer of kerosene can be poured on the water surface. If this latter method is used a density correction must be made to correct for the length of the kerosene column and thus obtain the true water level. This method cannot be used where there is a possibility that the kerosene can escape into the aquifer.

Volumetric and hydrometric methods are used to measure the volume of ground water produced from wells. To collect the requisite yield data may require the use of water-level recorders in conjunction with weirs and flumes, or flow-meters, the selection being based on proven reliability and utility as well as the expected range in flow.

Observations of ground-water temperature are commonly made by measuring the temperature of the water discharging from a well with a previously calibrated mercury thermometer. The usual accuracy of such thermometers is about 0.1° to 0.2° C or less. If down-hole temperatures are to be measured the use of thermistors is much more satisfactory.

Special thermistor devices designed in the United States have permitted an increase in the accuracy of temperature observations to within 0.001° C (Stallman, 1967) and are especially adaptable as remotely controlled devices. Much caution must be exercised when using such extremely sensitive equipment because the passage of the device through the well bore can disturb the temperature profile, as can internal convection currents. Commercial thermistor devices for measuring soil temperature within 0.01° C, down to a depth of 3 m, are available.

The electrical conductivity of ground water is related to the total dissolved solids content of the water. If curves relating conductivity to total dissolved solids are prepared for different types of water, conductivity readings can be used as a method of approximating the concentration of total dissolved solids.

Instruments designed for observing soil moisture are based on a variety of principles and are discussed in Section 7.4.5. The instrument currently in use is the neutron soil-moisture meter using radioactive plutonium or americium and beryllium as a source of radiation. Its weakness is related to low sensitivity in the near-saturation zone of the capillary fringe.

To satisfy requirements of durability and reliability under various climatic conditions, all instrument types should be evaluated and tested under a variety of climatic parameters, such as humidity and atmospheric pressure.

Standardization of individual instrument components, such as clock mechanisms, or relays, will permit mass production of hydrological measuring equipment. The various kinds of electronic equipment operating on the punched and magnetic tapes from water-level, water-quality and temperature recorders must be reliable. Such equipment must allow primary processing of information by computer—such as computing of 10-day or monthly sums, mean algebraic deviations, arithmetical means and probable observation errors; preparation of data tables; plotting of water-level graphs and piezometric maps; plotting of chemical-quality graphs and quality distribution maps.

7.4 Observation criteria

7.4.1 Objectives of a water-level measurement programme

7.4.1.1 General remarks

Observations of ground-water levels and the measurement of discharge of wells and springs are concerned with investigating: (a) patterns of annual and long-term change, i.e. the maxima and minima and the seasons in which they appear; the nature of these changes annually and over long periods; (b) dependence of water-level fluctuations and the discharge of flowing wells and springs on meteorological factors, on the stage of surface streams and reservoirs, and on resource development induced factors; (c) dependence of chemical composition and temperature of ground water on changes in the natural flow pattern; (d) directions of ground-water movement; (e) aquifer hydraulic characteristics; (f) areas of recharge and discharge.

Fluctuations in water level can be due to such diverse factors as changes in atmospheric pressure, loading of the land surface (e.g. passing trains), or seismic waves.

Fluctuations in ground-water levels are a measure of changes in ground-water storage which may take place under the influence of both natural and artificial factors (infiltration of precipitation, transpiration, evaporation, irrigation or pumping for water supply, etc.). The rate and magnitude of changes in ground-water levels are dependent upon both the extent and magnitude of the external influence, on the ground-water régime and on the permeability of the rocks in the aquifer and the zone of aeration. Therefore, observations of ground-water levels and discharge not only give direct information on the extent and rates of recharge to or discharge from the ground-water system but also permit determination of the hydrologic parameters of the water-bearing materials.

Ground-water systems respond rather slowly to external influences; thus there is no general need to register ground-water levels and discharge continuously. Under most conditions carefully planned intermittent measurements are sufficient.

To establish a rational frequency of measurements it is necessary to ascertain the pattern of ground-water replenishment or discharge, both under natural and disturbed conditions, taking into account the influence of natural and artificial factors and the variable hydrogeological conditions of the area under study.

For example, for granular sediments sharp fluctuations of water levels and outflow can be observed in some areas where the zone of aeration is 1 to 3 m thick; however, as the thickness of the zone of aeration increases, ground-water levels and outflow fluctuations become smoother and their magnitude decreases. Precipitation in an area of cavernous limestone may cause sharp fluctuations in ground-water outflow and of deep-seated (more than 20 to 30 m) water levels. To record rapid water-level fluctuations with adequate accuracy, frequent measurements or continuous recording may be required.

Rivers may become important sources of ground-water replenishment during flood periods. Recharge or damming of ground-water discharge by a rise in river stage causes a rise in ground-water levels in the zone adjacent to the river. The rate of rise can easily exceed 20 cm/day near the river. The width of the area affected depends on the river dimensions, the amplitude and duration of the rise in stage, and the hydraulic characteristics of the water-bearing formations. Commonly the rise in stage can effect water levels from several hundreds of metres to several kilometres on each side of the river. More frequent observations should be taken in the vicinity of the river during the flood period to enhance the accuracy of ground-water measurements throughout the entire affected zone. If hydrological parameters of the aquifer are to be determined from data derived during flood periods, continuous records of ground-water levels should be obtained by using automatic recorders.

Artificial factors can cause significant changes in conditions affecting ground-water recharge and discharge. For example, in irrigated areas pumpage can cause substantial water-level fluctuations within a day. The selected frequency of observation of water level or discharge of wells and springs must ensure that the principal fluctuations, as related to the natural and artificial factors responsible for the change, are recorded. Observations should be repeated on a recurrence interval which will minimize the effects of accidental errors in measurements.

7.4.1.2 Frequency of observations

The frequency with which ground-water levels or discharge of springs and wells are observed depends not only on the nature of the aquifer under investigation—especially on the extent and period over which these parameters change—but also on the problems that are to be solved.

At the beginning of investigations, when details of the ground-water system are not yet known, water levels and spring or well discharge measurements required for the dynamic analysis of the aquifer system are commonly made at regular intervals; for example, 10 times a month for water-table conditions and 5 times a month for artesian conditions at key observation sites.

The following method can be used to evaluate the possibility of reducing the number of observations: graphs of water-level fluctuations and discharge are constructed, according to several arbitrary frequencies of measurement, and compared. Thus, the same graph might be compiled three times by assuming that the number of observations had been only 10, 5 and 3 measurements a month. If the graphical error which arises from plotting the reduced number of measurements does not exceed the acceptable accuracy then the number of observations may be reduced.

The number of observations for artesian conditions may commonly be only 1 per month and for deep-seated water-table conditions perhaps only 3 per month. In areas bordering a river or lake, where the ground-water system is especially dynamic, registration of ground-water levels during flood periods should be carried out daily. Stage measurements of the adjacent river or lake should be of the same frequency and at the same times.

Daily records are also desirable during snowmelt periods in regions of shallow depth to ground water where rather sharp rises and declines of water levels (more than 1 to 2 cm/day) occur. Water-level extremes (maxima and minima), which are needed in calculating the ground-water balance, can be derived from daily observations. For similar reasons water levels in cavernous limestones should be measured daily during periods of possible ground-water recharge from precipitation. The pronounced water-level fluctuations characteristic of aquifers in karst areas are best documented by equipping the observation wells with continuous recorders.

On experimental balance plots for detailed study of infiltration, evapotranspiration and condensation, daily measurements are recommended, but during snowmelt, flood periods, storms and long-term rains, measurements 3 or 4 times a day are more desirable. It is preferable to equip wells and lysimeters with recorders during studies of this type.

The frequency of observations in regions where the ground-water régime is affected by artificial factors depends on the nature of those factors. In irrigated areas, the frequency of measurements must be related to changes in water applications for irrigation. If the areal variations in recharge to an aquifer are to be defined through observations in a group of wells in an irrigated area, then the readings should be taken accurately either several times a day or continuously. Continuous recording of ground- and surface-water levels is desirable when initiating drainage of water-logged areas.

Valuable data on hydrologic parameters can be obtained from continuous water-level records in areas where changes in large well operation have a significant effect on water levels. Ten measurements a month may be a minimum in a region of large wells but in a region of wells operating under a steady régime, the number of observations can easily be reduced to 1 per month.

To obtain static water levels using pumped wells the rate of recovery of the water-level after stopping the pump, should be measured and the static-level determined by extrapolation of the recovery data.

In areas of special detailed studies of the ground-water régime, as for example investigating the influence of transpiration, or changes in temperature, atmospheric pressure or seismic shocks on water-level fluctuations and yields of springs, the measurements should be continuously registered with automatic recorders.

Water-level measurements, regardless of the type of instruments used in their collection, should be referred to a fixed reference point (such as the edge of the well casing) whose altitude above sea level or any fixed datum has been determined by levelling. The accuracy of water-level measurements, made by such direct methods as a standard or electric measuring tape, should be ± 10 mm. Greater accuracy (± 1 mm) may be required when carrying out detailed investigations concerned with such effects as the influence of tides, atmospheric pressure, transpiration, condensation and tectonic movement in the earth on ground-water levels.

Volumetric methods are often used in measuring well or spring discharges in the range from 1 to 10 litre/sec. Stationary tanks can occasionally be used to measure larger discharges volumetrically. When measuring water discharge by volumetric methods a vessel of known capacity is placed under the discharge outlet and the rate of filling is timed. The discharge is determined according to the formula $Q = W/t$, where W is the capacity of the measuring vessel in litres and t is the time of filling in seconds measured by a stopwatch. Measuring vessels of a capacity such that the filling time is at least 30 to 40 seconds are recommended.

Spring discharges exceeding 1 to 5 litre/sec are more appropriately measured by means of thin-wall weirs which under properly controlled conditions have a high degree of accuracy. Weir-construction and measuring techniques are discussed in the WMO *Guide to Hydrometeorological Practices* (WMO, 1965).

Commercially produced flow-meters designed for installation in water-supply facilities can be used for measuring the discharges of production wells. Their design specifications and pertinent technical data are given in industrial literature. Flumes, weirs and orifices are also used for measuring larger discharges.

Mercury manometers and pressure gauges are used in obtaining head observations on flowing wells. The former provide relatively accurate measurements but are often inconvenient to handle in the field. Application of the latter should be with a stipulated permissible error of no more than 1.5 to 2.5 per cent of the full-scale range of the gauge.

When the head on a flowing well is not great, a tube can often be extended upward vertically to observe the head and head changes directly. These measurements can then be appropriately referred to the land surface.

The water-level measurements at a given location should be made at approximately the same time of day whenever possible. The observations of water-level and discharge are entered in the field registers which are kept in duplicate. Entries are made by an observer directly in the field; later the data is transferred to office records and the original field notes filed.

7.4.2 *Measurement of ground-water flow velocity*

The hydraulic conductivity (usually determined experimentally) of a porous rock material defines the rate at which a particular fluid moves through a unit cross-section of the material under unit pressure gradient at constant temperature. The fluid flows only through the pores of the material and, therefore, the hydraulic conductivity characterizes an average rate of flow through a number of capillary flow channels. Ground-water investigations may sometimes require the determination of the actual velocity of ground-water flow. The most widely applied method for determining the velocity of ground-water movement is based on the determination of hydraulic gradient and aquifer characteristics, from which the ground-water velocity can be calculated.

Other widely used methods of determining ground-water velocity involve tracers. In essence these methods require injection of a selected tracer into a well-located up gradient, in relation to the ground-water flow, from an observation well. If the direction of ground-water flow is not known several observation wells must be installed circumferentially around the injection well. The time of travel of the tracer and spacing between the injection and observation wells are approximately given by the formula:

$$t = \frac{r\theta}{KI} \quad \text{or} \quad r = \frac{KIt}{\theta} \qquad 7.4(1)$$

where r is the distance between the injection and observation wells in metres; K is the hydraulic conductivity of the rocks in metres per day; I is the hydraulic gradient describing the ground-water flow; θ is porosity of the rocks; and t is the period of time (days) between the injection of the tracer and its appearance in the observation well. Various dyes and radioactive isotopes have been used as tracers. In selecting tracers one should bear in mind that they should not: (a) precipitate on or be absorbed by the water-bearing rocks; (b) exist naturally in the waters under investigation; (c) decompose or lose their specific physical and chemical properties while percolating through the aquifer; (d) the chosen tracers must be easily detected.

Dyes have been most commonly used as tracers. These have included fluorescein, eosin, erythrosine and Congo red for alkaline waters and methylene blue, aniline blue and ponceaux red for acidic waters. Before injection the dye is usually dissolved in a small amount of either weak alkali or acid (approximately 100 to 150 g of dye per litre of solvent). The amount of dye required for the field velocity measurement depends on the permeability and thickness of the water-bearing material as well as the distance between the injection and observation wells. Table 7.4.2a provides an estimate of the amount of dye necessary.

TABLE 7.4.2a. Dye quantities used in measuring ground-water velocity in water-bearing rocks (in grammes of dry substance per 10 m of flow path)

	Clayey	Sandy	Fractured	Karstic
Fluorescein	5-20	2-10	2-20	2-10
Eosin	5-20	2-10	2-20	2-10
Erythrosine	10-40	10-30	10-40	10-40
Ponceaux red	10-40	10-30	10-40	10-40
Congo red	20-80	20-60	20-80	20-80
Methylene blue	20-80	20-60	20-80	20-80
Aniline blue	20-80	20-70	20-80	20-80

In using dye tracers, water samples are taken from the observation wells at intervals appropriate to the anticipated time of travel. The interval can be varied to suit the local situation. Considerable care must be exercised to prevent disturbing the natural flow pattern by the sampling. Under some conditions the degree of colouring of the sampled water can be determined visually by reference to previously established colorimetric standards. If the field conditions cause high dilution rates, instrumental detection may be necessary. When using fluorescein or other fluorescent tracers the use of a fluorimeter gives a highly sensitive detection device. The dye concentration at the moment of injection is considered to be the 100 per cent concentration reference (Dunn, 1957). Pieces of activated carbon can be suspended in the water in the observation well in order to concentrate the fluorescein and thus facilitate detection.

Dyes which have received wide use in the United States for time-of-travel studies of rivers are the rhodamine series. Several of these could be used in ground-water tracer studies and their detectability limit is a few parts per billion using a fluorimeter as a dilution device. Table 7.4.2b lists these dyes and their properties.

TABLE 7.4.2b. Rhodamine dyes used as tracers in time-of-travel studies

Product	Specific gravity	pH or acidity	Comments
Du Pont Rhodamine WT Solution 20%	1.19 ± 0.02 at 20/20° C	8.5 ± 0.5 at 20° C	This colour was developed to exhibit high tinctorial strength and a low tendency to stain silt, dirt and other suspended matter in shallow and inland waters.
Du Pont Rhodamine BMG Solution 40%	1.02 to 1.04 20/4° C	9.0 ± 0.5	Represents an improvement in aqueous solubility over Du Pont Rhodamine BA Solution 40%. This was accomplished by substituting ethylene glycol for the acetic acid in the formula. This product also eliminates the hazard of handling acetic acid.
Du Pont Rhodamine BA Solution 40%	1.030 + 0.005 20/4° C	30.0 ± 1.0 % expressed as acetic acid	Formulated to the specifications of Johns Hopkins University. This product exhibits less acidity and more tinctorial strength than Du Pont Rhodamine BA Solution 30%.
Du Pont Rhodamine BA Solution 30%	1.030 ± .005 20/4° C	37.5 ± 1.0% expressed as acetic acid	Developed specifically for Columbia University to give the same specific gravity as sea water.
Du Pont Rhodamine B Solution	1.120 ± 0.005 at 20/4° C	48.0 ± 1.5% expressed as acetic acid	250 : 100 versus Du Pont Rhodamine B Extra Standard (Spectrophotometric).
Du Pont Rhodamine TLX-670 Solution	1.00 to 1.02 at 25/4° C	9.0 ± 0.5 at 25° C	This solution was developed as a low-acidity type.

Salt solutions are commonly used as the tracers. The natural salt content of water in the aquifer should be predetermined by collecting and analysing samples from the observation wells prior to injection of the tracer. The most commonly used salts are: sodium chloride (table salt), lithium chloride, ammonium chloride and calcium chloride. Another chemical tracer is potassium dichromate. Approximate doses of the salt tracers are given in Table 7.4.2c.

The concentration of most salt tracers can be determined in the observation well by measuring the electrical conductivity of the water which flows through the well. For this purpose special electrodes insulated from the well casing are lowered to a position opposite the screen. These electrodes are connected to a battery, an ammeter and a rheostat. By measuring the current in this circuit it is possible to determine the electrical conductivity of the water moving through the observation well.

7.4.2c. Recommended doses of salt tracers for injection into well

Salt	Dose (in kg)	Distance between injection and observation wells (in metres)
Sodium chloride	10-15	5-7
Calcium chloride	5-10	3-5
Ammonium chloride	3-5	2-5
Lithium chloride	0.010-0.015	2-5

During recent years methods for determining ground-water flow direction and velocity by means of radioactive tracers have been developed (Ferronsky, *et al.*, 1968; IAEA 1968). The isotopes selected as tracers must satisfy the following requirements: (a) energy of radiation must permit detection of the tracer in the well or in the collected samples; (b) the half-life of the isotope should be sufficiently long to permit detection; and (c) the sorption losses of the isotope in the environment porous medium must be negligible. Unfortunately, the injection of radioisotopes into a ground-water system presents a health hazard and the quantities of tracer used must be limited so that the maximum permissible human consumption levels are not exceeded. Tritium, I-131 in the form of sodium iodide or potassium iodide, S-35 in the form of sodium sulphate and other isotopes have been recommended by a number of researchers for use as radioactive tracers (Dimaksyan, 1966; IAEA, 1968). The introduction of large volumes of tritium into the hydrological cycle from nuclear tests can be utilized for tracing ground-water movement.

Radioactive tracers, as well as salts, are used to detect vertical movement of water in wells both while they are being pumped and under static conditions. These techniques are especially valuable in multi-aquifer systems or in areas where the dominant permeability is secondary. Intra-aquifer movement of water and zones of highest yield can easily be determined using tracer methods. The use of tracers is most useful in measuring very low vertical velocities.

A number of radioactive-tracer techniques are available for determining direction and velocity of ground-water flow. Some recommendations for the use of radioactive isotopes, determination of the distance between the injection and observation wells, choice of tracers and methods for processing the observation data are described in several guides (Ferronsky *et al.*, 1968; IAEA, 1968).

From the results of determining ground-water velocity by means of tracers (dyes, salts or radioactive isotopes) a graph showing changes in tracer concentration with time can be constructed, similar to that shown in Figure 7.4.2. Depending on the tracer used, the vertical axis corresponds to the concentration of the tracer detected in the observation well or the percentage of dye or radioactive isotope, or current in amperes (or resistance in ohms); the horizontal axis is for the co-ordinate of time. Point A on the graph (Fig. 7.4.2) corresponds to the appearance of the tracer in the observation well and point B is the moment of maximum concentration. The ground-water flow velocity is determined by dividing the distance from the injection well to the observation well by the elapsed time from the moment of tracer injection to the moment of arrival of its maximum concentration in the observation well.

Aside from technical problems associated with any tracer test method, the principal problem in applying the results is one associated with the size of the aquifer sample. In general, most tracer tests only investigate a very small segment of the aquifer and how representative of the whole system this small segment may be is problematical. Tracer test methods may supply specific answers for small areas but indirect methods, using aquifer characteristics and regional gradients, should be used wherever possible when attempting to obtain velocity of ground-water flow over large areas.

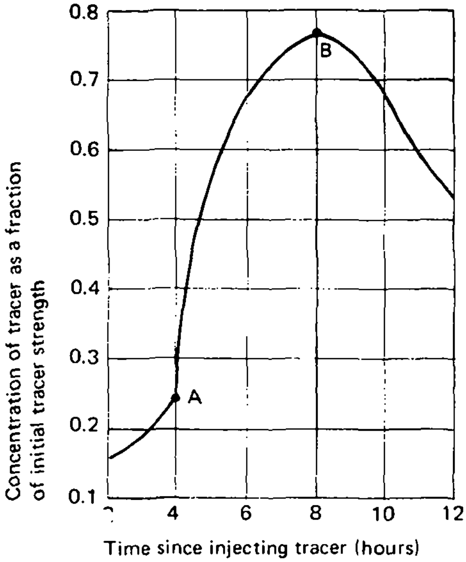


FIG. 7.4.2. Graph of change in tracer concentrations in observation well.

The following geophysical methods for determining ground-water flow velocity have been applied while carrying out ground-water investigations: the self-potential field method and the electrical resistivity method. Description of these methods can be found in special publications (Legette, 1950; Matveev, 1963; Jones and Skibitzke, 1956). One method of investigation requires the injection of a concentrated solution of sodium chloride into a well. An electrode connected to a battery is then submerged in the well. Then eight equally spaced (45° intervals) radials originating at the well and extending outward approximately three times the depth to the aquifer are marked on the land surface.

A detection electrode is then moved along each radial while measuring the potential difference by means of conventional potentiometers. Graphs of the data are constructed, from which potential lines are drawn. If this procedure is repeated at different time intervals, the equipotential lines protrude in the direction of the ground-water flow, due to the flow of the sodium chloride used as the electrolyte. This makes possible the determination of the direction and velocity of ground-water flow.

The resistivity method, based on the dilution of electrolyte within a well due to ground-water flow can also be applied in special cases. A concentrated solution of electrolyte (usually sodium chloride) is injected into the well and carefully mixed. Before and after injection periodic (in equal time intervals) measurements of water resistance are made by means of a resistivity meter. The rate of the electrolyte dilution, as well as the ground-water flow velocity, can be determined from the changes in resistivity.

One of the important advantages of these geophysical methods lies in the possibilities they offer for determining the ground-water flow velocity if only one well is available. However, the effect of the borehole itself upon the lines of ground-water flow cannot always be neglected, particularly where the well casing may be inadequately slotted.

7.4.3 *Observations of ground-water temperature*

Ground-water temperature is one of the parameters offering a means for observing changes in the state of the ground-water régime in time and space, which relate to conditions for recharge as well as the influence of various natural and man-induced factors on the régime. Because ground water saturates much of the rock materials in the upper layers of the earth's crust the water temperature reflects in part the temperature of the water-bearing rocks. The mobility and thermal capacity of ground waters, however, serves to redistribute some of the heat within the stratosphere and to influence the development of the geothermal régime within this sphere. Thus the ground-water temperature régime cannot be considered isolated from the temperature régime of the rocks comprising both aquifers and the adjoining impervious formations.

Observations of the temperature régime in a section penetrated by wells afford not only a view of ground-water temperature changes in time and space but also contribute to the solution of a number of theoretical and practical problems, such as follows:

1. Establishing the regional features of the hydrogeothermal régime and defining (a) the thermal zones identifiable with the zone of aeration, and unconfined and artesian ground waters; and (b) determining the depths at which thermal waters are confined throughout the region.

   To obtain data for solving this class of problem a key thermometric network should be established. For this purpose the deepest available wells should be chosen to encompass all the significant geological-hydrogeological and physico-geographical features in the region.
2. Analysis of specific ground-water problems with respect to the following features: (a) detection of the major permeability changes in a geological section penetrated by a well; (b) definition of conditions favouring interaction between surface and ground waters; (c) identifying the location and extent of potential sources of ground-water discharge; (d) assessing the thickness and intensity of karst development; (e) forecasting the expansion of cones of depression at water-development sites and the conditions relating to water leakage between aquifers; (f) determining the thermal balance to show the quantity of heat removed from the earth's interior by ground water.

Solutions to the foregoing class of problem (types 2(a) to 2(f)) require a network of wells and supplemental work in accordance with the specific tasks.

Most of the enumerated problems can be analysed by hydrogeothermometric methods if it is recognized that the old concepts concerning the damping in amplitude of temperature fluctuations down to depths of 20 to 40 m (of heat transfer from the earth's surface by conductivity) and the existence below this depth of the so-called zone of constant temperature require further clarification. Results of special hydrogeothermal investigations have shown that the penetration into the earth of the ambient temperature effect to about the foregoing limiting depths (closed type of hydrogeothermal zone, (Fig. 7.4.3,I) is only possible if the geologic section is characterized by very thick formations of impermeable rocks and the region is characterized by very weak erosional relief. If aquifers are present in the vertical section. so that water and heat exchange occur, and the intensity of this exchange is controlled principally by the erosional down-cutting, the physical properties of the rocks and the rate of recharge, then the temperature régime and the pattern of temperature zones become far more complex.

The open type of zone (Fig. 7.4.3,II) is typical of areas underlain by a layer of fairly permeable rock of about 30 to 50 m in thickness. The combined type of zone (Fig. 7.4.3,III) is found in areas underlain by formations of alternating permeable and impermeable rocks down to the base level of erosion with an impervious bed more than 25 to 30 m thick in the upper part of the sequence.

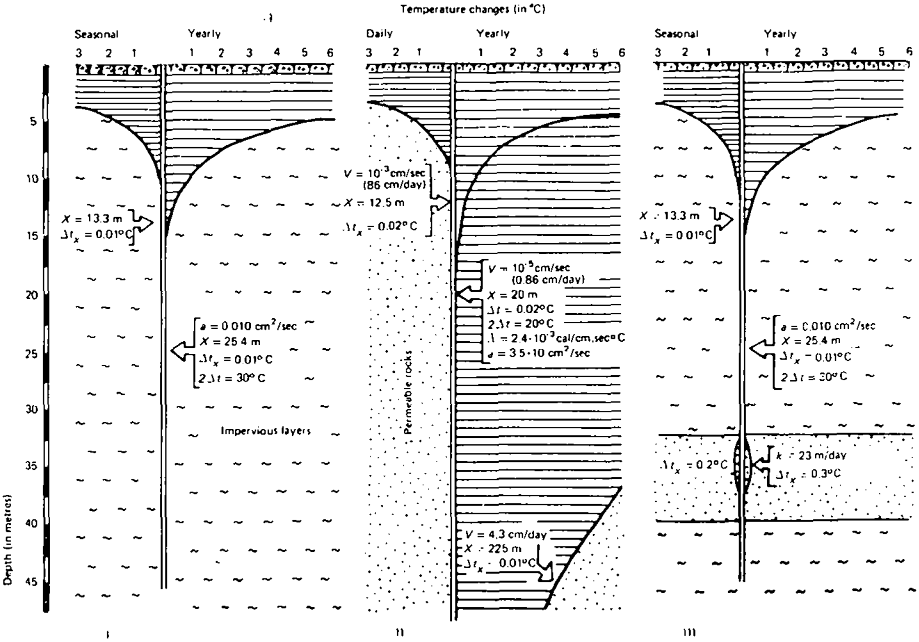
Selection of observation points for collecting temperature data is dictated by the particular problems to be analysed and by the hydrogeothermal investigational techniques proposed for their solution. Thus if it is desired to determine the limits of the zone in which water actively circulates in a certain region the observations of the temperature régime must embrace the entire diversity of geological and hydrogeological conditions throughout that region. This requirement stems from the fact that the temperature régime for the zone of active water circulation is greatly dependent upon exogenic factors and upon the features that influence the distribution of solar heat absorbed by the surface of the earth. The number of observation points required for the solution of particular problems must usually be separately determined for each case. When studying the relationship between ground and surface waters the thermometric observations must be extended to the surface-water reservoirs and any sites of ground-water discharge.

Studies of the ground-water temperature régime and the water-bearing rocks demand high accuracy of measurement—at least to within ± 0.01° C. But such accuracy is not needed if the ground-water temperature régime is being studied from the viewpoint of its practical development and use. The accuracy may then be one order lower.

The scheduling of observations depends upon the problems being analysed and the specific geological and hydrogeological conditions in the study area. Temperature measurements in wells maintained as part of a key regional thermometric network might be scheduled as follows: (a) four times on a selected day once a month during the first year (at 1, 7, 13 and 19 h); (b) four times on a selected day each three-month period thereafter.

The time interval between measurements should obviously be adjusted to suit the particular problem; thus, it might be shortened considerably for wells experiencing sudden fluctuations in temperature.

Temperature measurements should be collected through the full depth range of wells in the key network regardless of the contents of the well (water or air). The interval between temperature readings down to a depth of 40 m is recommended as no more than 2.5 m (2.5, 5.0, 7.5 m, etc.); below this depth readings are recommended at the 50,



x = limiting depth of ambient temperature effect.
Δt_x = magnitude of ambient temperature effect at depth x.
$2\Delta t$ = temperature range at land surface.
a = heat conduction in rock in cm²/sec (analogous to transmissivity).
λ = thermal conductivity of rock in cal/cm.sec.°C.
V = rate of infiltration in cm/sec. (cm/day).
K = hydraulic conductivity in m/day.

FIG. 7.4.3. Principal types of hydrogeothermal zones (A: closed; B: open; C: combined) (after Frolov, 1966).

75, 100, 125, 150, 200, 250, 300, 350, 400 m depth levels; and below this point readings are recommended only at 100 m intervals. Fixed points for temperature measurements may be established in each well opposite the principal aquifer(s) and at a distance of 0.5-2.5 m on each side of the aquifer contact(s) with impervious beds.

Depending upon the water-circulation conditions, the observations of the temperature régime may require days or decades to reveal significant information on water movement at various depths (Frolov, 1966, 1968). The zone of aeration and the upper part of an aquifer, where infiltrating water mixes with the aquifer water, are distinguished on graphs of the temperature régime by the larger values of temperature change. In the saturated zone, the temperature changes at various levels depend on the ground-water flow velocity and increase with increased velocity. Furthermore, the greater the flow velocity the less time required to observe the temperature changes. This permits a layer-by-layer hydrodynamic separation of rock sequences under conditions that otherwise might appear visually homogeneous.

7.4.4 *Observations of ground-water quality*

Regional variations in ground-water quality can be determined by sampling water from wells drilled at sites intended to give representative coverage of the various conditions of occurrence. Partial analyses to determine the concentrations of the principal chemical constituents in a water may provide sufficient data for many investigations. Such analyses are made, for example, of water samples collected during the drilling of wells, for water-balance studies and for experiments (using three or more wells) designed to calculate the rate of ground-water flow. The constituents commonly determined are shown in Table 7.4.4.1. Soil and rock samples are also collected to determine the amounts of water-soluble salts present in the zone of aeration.

More detailed studies require the collection of water samples at regular intervals from selected observation wells and soil samples from test borings near the observation wells. The results of the partial chemical analyses of water, water extracts and possibly of soil solutions obtained by compressing the samples are useful in determining the balance of water-soluble salts in the areas under investigation.

In addition to the collection of ground-water samples from observation wells, samples are collected for chemical analysis of the water in surface streams and other water bodies such as lakes or reservoirs whose hydraulic connexion with the ground-water régime needs to be examined.

After ascertaining the chemical composition of ground water from various aquifers (by means of partial analyses of water samples), waters of distinctive chemical composition are identified and areas selected in the basin which appear most promising for practical and scientific studies. With this information, and bearing in mind the requirements of the over-all investigation, representative observation wells are selected to enable a study to be made of the changes in chemical composition of the ground-water. Additional chemical constituents requiring periodic determination are itemized. Wells are selected from which water samples can be collected for more detailed chemical analysis.

Lines of wells along the principal directions of ground-water flow are usually chosen to study regional variations of the chemical composition of the water. In each major geomorphological unit (such as plateau, high and low structural and depositional terraces, etc.), one or two reference wells are selected with screens installed in the principal aquifers that are significant as sources of water for practical use and that support base-flow in the streams. If certain formations only yield water seasonally, or certain minor aquifers are hydraulically connected with the principal aquifers, groups of observation wells with screens placed in these seasonal or minor units are arranged near the reference wells.

During the first stage of hydrochemical investigation, water samples from the principal aquifers are taken on a seasonal basis, the first samples being subject to complete analysis. Later samples are given only partial analysis.

7.4.4.1 Types of analyses

The type of water analysis undertaken depends upon the objective of the investigation Thus a water might be assessed with respect to its suitability for drinking, irrigation and industrial purposes, or to its aggressiveness to concrete. Not uncommonly it is necessary for the solution of specific problems to know the content of all major constituents. An estimate of the total concentration of dissolved ions can be obtained by determining the specific electrical conductance of the water.

The specific conductance of an electrolyte is the reciprocal of specific resistance and is expressed in mhos (reciprocal ohms). Specific resistance is the resistance, in ohms, of a column of solution 1 cm long and 1 cm^2 in cross-section. In most waters, the conductance is so low that the micromho, that is, a reciprocal megohm, is used as the unit of expression. The specific conductance is a measure of the ability of the water to carry an electric current and is therefore an indication, within rather wide limits, of the total dissolved solids of the solution.

For most natural waters of mixed type, the specific conductance, in micromhos,

TABLE 7.4.4.1. Principal features of several types of water-quality analyses

Type of analysis	Items to be determined	Features of the analysis
I	Physical properties; conductance, hardness, Cl^-, SO_4^{--}, HCO_3^-, CO_3^{--}, pH. Qualitative determination of Fe^{++} + Fe^{+++}; NO_3^-; NO_2^-; NH_4^+.	*Partial analysis* made to give preliminary information about the waters of an area. May be accomplished in the field by means of simplified methods. The sum of Na^+ and K^+ is calculated by difference. Results of the analysis cannot be checked for accuracy.
II	Physical properties; conductance, Cl^- SO_4^{--}; HCO_3^-; CO_3^{--}; NO_3^-; Ca^{++}; Mg^{++}; Fe^{++}; Fe^{+++}; pH; CO_2 (free); dry residue; NO_2^-; NH_4^+; redox potential.	*Partial analysis with added detail* gives general water characteristics. This analysis can be made in fixed or well-equipped mobile laboratories. An approximate check of accuracy of the analysis can be made by comparing total ionic content with weight of the dry residue. The sum of K^+ and Na^+ is calculated by difference.
III	Physical properties; conductance, Cl^-; SO_4^{--}; HCO_3^-; CO_3^{--}; NO_3^-; NO_2^-; F^-; Ca^{++}; Mg^{++}; Na^+; K^+; Fe^{++}; Fe^{+++}; Mn^{++}; NH_4^+; R_2O_3; SiO_2; pH; CO_2 (free and total); H_2S; dry residue; redox potential.	*Complete analysis* for representative waters or waters of specific interest. This analysis is generally accomplished in fixed laboratories. The analytical results enable a check of the analysis both from the weight of the dry residue and by comparing total amounts of milligramme equivalents of anions with cations.
IV	Selected physical properties for special purposes.	*Special analysis* including determinations which are not listed in the foregoing types, or partial analyses with determinations of a few additional components only. Choice of determinations depends on the purpose of the analysis.

multiplied by a factor of 0.65 $\pm$ 0.1 approximates the residue on evaporation in milligrammes per litre. This does not approach an exact relation because the conductance of a solution is dependent on the type and total quantity of ions in solution. More precise relations can be developed for specific water types. The factor of 0.65 is applicable only with comparatively dilute solutions and usually increases as the total dissolved-salt content exceeds 2,000 to 3,000 mg/l. For waters that contain appreciable concentrations of free acid, caustic alkalinity or sodium chloride the factor may be much less than 0.65. The factor for some other specific types of water may be higher. Non-ionized silica will also disturb the ratio of residue on evaporation to conductance. With similar limitations, the specific conductance divided by 100 approximates the milliequivalents per litre (Sect. 7.4.4.4.) of anions or cations. This relation is particularly helpful in detecting the location of error (in anions or cations) as well as for estimating the comprehensiveness of an analysis.

One might have a rough judgement of the water composition after determining the total hardness; the concentrations of HCO_3^-, CO_3^{--}, Cl^- and SO_4^{--}; and the pH. Some natural waters contain considerable amounts of NO_3^- ions and noticeable quantities of iron. Analysis for the preceding constituents is usually supplemented, therefore, with quantitative determinations of NO_3^-, Fe^{++} and Fe^{+++}.

To obtain a more detailed indication of the ionic composition, the concentrations of Ca^{++}, Mg^{++}, Na^+ and K^+ ions, the dry residue, the sum of sesquioxides (R_2O_3) and silicic acid (SiO_2) are separately determined. Total hardness need not then be determined. When appraising the aggressiveness of water with respect to concrete, free and total carbon dioxide are determined. When analysing water intended for drinking, F^-, NO_2^- and NH_4^+ ions, redox potential and such physical properties as colour, odour and clarity are determined. A summary of the main features of various types of water analyses is given in Table 7.4.4.1.

A considerable amount of information about the composition of water can be obtained from analyses made in the field at the time of sampling. The content of such volatile or unstable components as NO_2^-, NH_3, free CO_2, H_2S, free O_2, Fe^{++} and pH can be determined satisfactorily only in the field or with special sample preparation. However, some of the simple analytical methods have a low accuracy, and care should be taken to ensure that errors do not obscure the natural fluctuations in quantities of unstable constituents.

7.4.4.2 Pore solutions

The salts or ions which give rise directly to the chemical constituents occurring in natural waters are present in rocks either in the solid form, as salts or adsorbed ions, or in solution in the pores of the rock. These constituents have been derived primarily from the chemical weathering of igneous and metamorphic rocks.

In a detailed hydrochemical study the composition should be determined of the salts and adsorbed ions present in the rocks and capable of contributing to the chemical composition of subsurface waters. The chemical nature of waters in aquifers can be determined by sampling the ground water, but in some cases it may be necessary to determine the nature of pore waters in clays or in rocks not fully saturated, as for example in the zone of aeration or in soils.

Water extracts give a general indication of the amount and approximate composition of soluble salts in rocks. The results obtained, however, are dependent upon the method used to prepare the water extract.

Pore solutions can sometimes be obtained for analysis by compressing the rock or the solutions may be displaced by non-aqueous solvents. Sometimes both techniques are used together. The samples obtained in this way may give a better idea of qualitative and quantitative compositions of natural solutions than samples of water extracts.

However, neither compression techniques nor water extracts reliably indicate the composition of water-soluble salts in rocks or the content of adsorbed ions.

The following constituents are usually determined when analysing water extracts: pH, dry residue, $Na^+ + K^+$, Ca^{++}, Mg^{++}, Cl^-, SO_4^{--}, HCO_3^- and CO_3^{--}. Sometimes the nitrogen compounds (NO_2^-, NO_3^-, NH_4^+) or I^-, Br^-, etc., are determined. Two components, SO_4^{--} and CO_3^{--} can be determined by treating solid samples with hydrochloric acid.

7.4.4.3 Methods of sampling

To accomplish steps 2, 3, and 4 described near the beginning of Section 5.6.1, water samples for an expanded type of partial analysis (analysis type II in Table 7.4.4.1) are taken seasonally from the reference hydrochemical observation wells. In water-balance studies, samples may be required each month during the first two years. Preliminary-type partial analyses (type I in Table 7.4.4.1) are made at the outset on water samples taken while drilling observation wells.

When studying salt balances, in addition to sampling water from various sources for chemical analysis, soil and rock samples from the zone of aeration are taken at representative locations in the basin, and chemical analyses are made of water extracts. The first sampling is undertaken when observation wells are being drilled and subsequent (once each season or quarterly) samplings are made from specially-bored exploratory holes near observation-well sites. When the water table is less than 4 m below the land surface, soil and rock samples are taken at vertical intervals of 10 cm throughout the zone of aeration. This may be done in each of three exploratory holes drilled within 3 to 5 m of the observation well and spaced 1 or 2 m apart.

Where the water table is at a greater depth and where the zone of aeration is of homogeneous composition samples are taken at vertical intervals of 10 cm within the first 1.5 to 2 m below the land surface and also within the capillary fringe. In the intermediate subzone, samples are taken at 0.5 to 1 m vertical intervals. When the soil and rock samples are being collected for chemical analysis a duplicate set is taken from the same depths and sealed in suitable glass containers for subsequent determination of the natural soil-moisture content. From one of the exploratory holes, water samples are collected to determine the chemical composition of water near the water table.

The most suitable times for collecting soil and subsurface water samples are when the water table reaches its minimum, maximum and average-annual positions. This enables water and salt balances to be calculated for the various seasons of the year.

The following weights are recommended for soil samples for the determinations indicated:

300 g: chemical composition of a water extract (samples are put into small sacks or plastic bags).

50 g: soil-moisture content (samples are put into air-tight glass containers).

500 g (min.): composition of pore waters in soils and rocks by compression (samples are put into glass containers to preserve the natural moisture content).

The volumes of water samples required for chemical analysis depend on the type of analysis to be made. Recommendations are as follows:

0.5 litre: partial analysis, field, type I.

1 litre: partial analysis, expanded, type II.

2-5 litres: complete analysis, type III, including determination of some minor constituents.

The volume of a sample for spectral analysis of minor constituents must be sufficient to ensure a dry residue after evaporation of not less than 60 mg. The volume of samples for determining Mo, Zn, or Cu should be 0.3 to 0.5 litre, and for determining Pb should be 0.8 to 1.0 litre.

The following precautions should be taken when collecting water samples for chemical analyses.

1. If possible, samples should be obtained so that the escape of dissolved gases (free CO_2) and oxidation (for instance, conversion of Fe^{++} into Fe^{+++}) is prevented. The samples must be collected as the water emerges from the ground or from the centre of the flow at the well head.
2. Before collecting water samples from any excavations or wells, stagnant water should be pumped out as it may have a composition different from that of ground water *in situ*. In polluted water, sulphate reduction can occur which would affect the concentration of SO_4^{--}, free CO_2, pH and possibly other constituents. The volume of water pumped before collecting water samples should amount to 1.5 to 2 times the volume of water in the excavation or well.
3. When collecting water samples from a group of excavations, the shallowest should be pumped and sampled first.
4. Ideally, water samples should be taken from wells at the depths of the screens by means of a special depth sampler.
5. When collecting a water sample, enough water should be passed through the bottle by means of a tube lowered to the bottom of the bottle to renew the water in the bottle two or three times. This requirement is mandatory if the sample is intended for determination of gases (e.g. free carbon dioxide).

As some dissolved constituents (Ca^{++}, Mg^{++}, Fe^{++}, Fe^{+++}) may precipitate from a water sample during storage under laboratory conditions the analyses should be completed as soon as possible after the samples have been collected. Rock particles should be excluded from water samples, and it should be recognized that samples stored in glass vessels can gain or lose minor constituents. If a determination of such constituents is to be made, the sample must be collected in a specially prepared container and preserved at the time of collection, usually by acidification.

7.4.4.4 Representation of chemical data

The preliminary examination of chemical data is aimed at classifying the chemical analyses of water and soil solutions, identifying hydrochemical types of waters and describing changes in the chemical composition in time and space. Compilation of hydrochemical maps, determination of the chemical and biological reactions affecting the composition of the water, establishment of the sources of the solutes, correlation of the hydrochemical régime with other factors and calculation of salt balances are undertaken during the final processing of the data.

The preliminary classification of the chemical analyses should be based on such general features as the change in concentration of certain chemical constituents in relation to water movement and circulation. The length of time the water is in contact with the rock may be related to the total ionic concentration.

The regional distribution of the six principal components (HCO_3^-, SO_4^{--}, Cl^-, Ca^{++}, Mg^{++}, $Na^+ + K^+$) should be determined first. Brodsky (1953) proposed a method of displaying the analytical results (Fig. 7.4.4.4a) in which the concentration of dissolved ions determined in each analysis is designated by a symbol which is entered in the appropriate square of the graph. Grouping of the analyses shown on the classification diagram is made after taking into consideration the geological and hydrological situation. Such groupings may be represented on the diagram by means of a colour code. Thus the principal features of the spatial variations in chemical composition of waters and the general processes leading to the observed composition are illustrated.

When comparing the ranges in concentration of individual chemical constituents in a large number of water samples, collected within an extensive area or over a long period of time, it is desirable to show the results of many analyses on a single graph. The analytical results should be expressed and plotted preferably in milliequivalents per

		HCO_3		SO_4		Cl	
		Cl	SO_4	HCO_3	Cl	SO_4	HCO_3
Ca	Mg						
	Na		○	△			
Na	Ca				□		
	Mg						
Mg	Ca						
	Na					◊	

○ Dry residue, up to 1 g/1.
△ Dry residue, from 1 to 5 g/1.
□ Dry residue, from 5 to 30 g/1.
◊ Dry residue, more than 30 g/1.

FIG. 7.4.4.4a. Classification diagram for illustrating chemical composition of ground water.

litre, rather than in percentages, as this enables each component to be viewed and compared in terms of its absolute concentration.

For general comparison of the chemical composition of a number of water samples a semi-logarithmic graph, such as the type proposed by Schoeller (1956, 1962) can be used to show simultaneously the absolute concentration values and their relative magnitudes (Fig. 7.4.4.4b). The abscissa (arithmetic scale) provides a convenient base along which the principal chemical constituents Ca, Mg, Na + K, Cl, SO_4, HCO_3 + CO_3 are arbitrarily and evenly spaced from left to right (Fig. 7.4.4.4b). The ordinate (logarithmic scale) refers to the concentration of the selected ion in milliequivalents per litre. The milliequivalent value is calculated by dividing the weight of the constituent in milligrammes by its atomic weight, divided in turn by its valence. For example, the weight of the constituent Ca is divided by 20; Mg by 12, Na by 23, K by 39, Cl by 35.5, SO_4 by 48, HCO_3 by 61 and CO_3 by 30. Each of the curves (Fig. 7.4.4.4.b) represents the analysis of a single water sample. The relationship between two chemical constituents in two different water samples is expressed comparatively by the slopes of the straight lines connecting these constituents.

Parallel lines on the semi-logarithmic graph express identical relationships. For example, if the line joining rSO_4' and rCl' of one water and the line joining rSO_4'' and rCl'' of another water are parallel, the ratios of these elements are the same: $rSO_4'/rCl' = rSO_4''/rCl''$.

With this type of graph it is also possible to determine, from the inclination of the connecting lines, the ratios of the elements in a given type of water to each other, more particularly those elements which are deemed characteristic or of prime importance, rMg/rCa, rSO_4/rCl, $(rCl\text{-}rNa)/rCl$, rNa/rCa and rNa/rCa.

It is also possible to find out whether the solubility products have been reached. For instance, for the product $[SO_4]$. $[Ca]$, all that is needed is to join the points rSO_4 and rCa by a straight line; if the straight line cuts the vertical $CaSO_4$ (located half way between rSO_4 and rCa) below the point of saturation S, saturation has not been reached. On the other hand if the point of intersection is above S, there is supersaturation. The location of the point S depends on the ionic strength, as indicated on scale I.

Solubility of $CaSO_4, 2H_2O$

$CaSO_4$ (milliequivalent)		Ionic strength		Values to be added to S'o of $CaSO_4$ for various temperatures, at $\mu = 0$	
At 14°	At 20°	At 14°	At 20°	Temperature	S'o
25	31	0.050	0.062	0°	−3.8
34	40	0.118	0.130	10°	−1.3
41	46	0.182	0.192	20°	0
50	55	0.300	0.210	30°	1.2
54	59	0.358	0.368	40°	1.6
65	69	0.650	0.638	55°	1.0
84	88	1.168	1.176	60°	0
97	101	1.694	1.702		
101	105	1.952	1.960		
104	107	2.208	2.214		
106	107	2.462	2.464		
104	105	2.708	2.710		
103	103	2.956	2.958		
100	100	3.200	3.200		
93	93	3.686	3.686		
87	87	4.174	4.174		
81	81	4.662	4.662		
78	78	5.156	5.156		

It is also possible to find out whether water will dissolve or deposit limestone. The equilibrium pH of a water is given by the equation:

$$\text{pH eq.} = \log \frac{Kc'}{K_2''} - \log [HCO_3] - \log [Ca^{++}]$$

This being so, one can plot on a scale midway between $rHCO_3$ and rCa the equilibrium points corresponding to the various pH eq. values. Thus if the line drawn between $rHCO_3$ and rCa crosses the equilibrium scale at a pH higher than the real pH of the water, the latter will dissolve limestone, while if it crosses lower down there will be supersaturation.

The position of the equilibrium pH is directly related to the ionic strength μ of the water.[1] These pH values of the water must therefore be moved upwards by an amount equal to that indicated at the bottom of scale 2.

Account must also be taken of the temperature of the water and the equilibrium pH moved down by an amount equal to that indicated in the table. These corrections correspond to the values shown in the table below (7.4 page 19).

The straight line joining $rHCO_3$ to rCa intersects the perpendicular drawn one-third of the way from $rHCO_3$ and two-thirds of the way from rCa at a point K_r giving the equilibrium pressure δ of the CO_2. This gives:

$$K_r = \sqrt[3]{\frac{a\delta}{22.4} \frac{K_2''}{K_1''K'_c}} \times 10^3$$

a being the coefficient of dissolution of carbonic acid gas in the water. The position of k_r also varies directly with the ionic strength μ as indicated on scale 3 in the Fig.7.4.4.4c.

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[1] *The ionic strength of the water may be calculated using milliequivalents on the basis of the following equation:*

$$\mu = 0.001 \left[\frac{rCl + rHCO_3 + rNO_3 + rNa + rK}{2} + rSO_4 + rCO_3 + rCa + rMg\right]$$

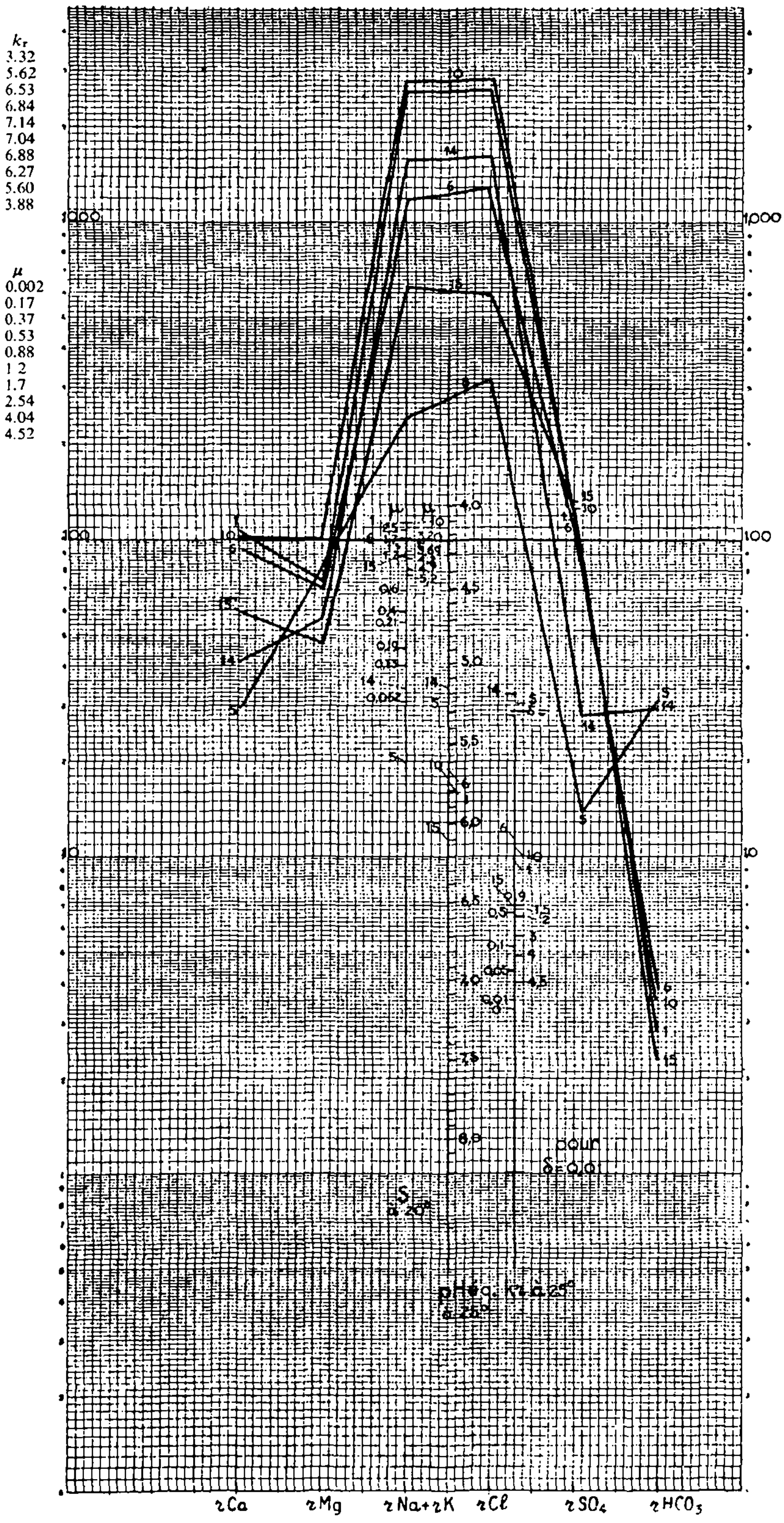


FIG. 7.4.4b. Example of semi-logarithmic diagram ('Waters of West Texas', *Analyses in Problems of Petroleum Geology*, 1934, p. 883-4).
S: scale of products (rSO_4) (rCa) for various ionic strengths at 20°. pH eq.: scale of equilibrium pH of solubility of $CaCO_3$ at 25°.
k_r: scale of k_r at 25° for various ionic strengths.

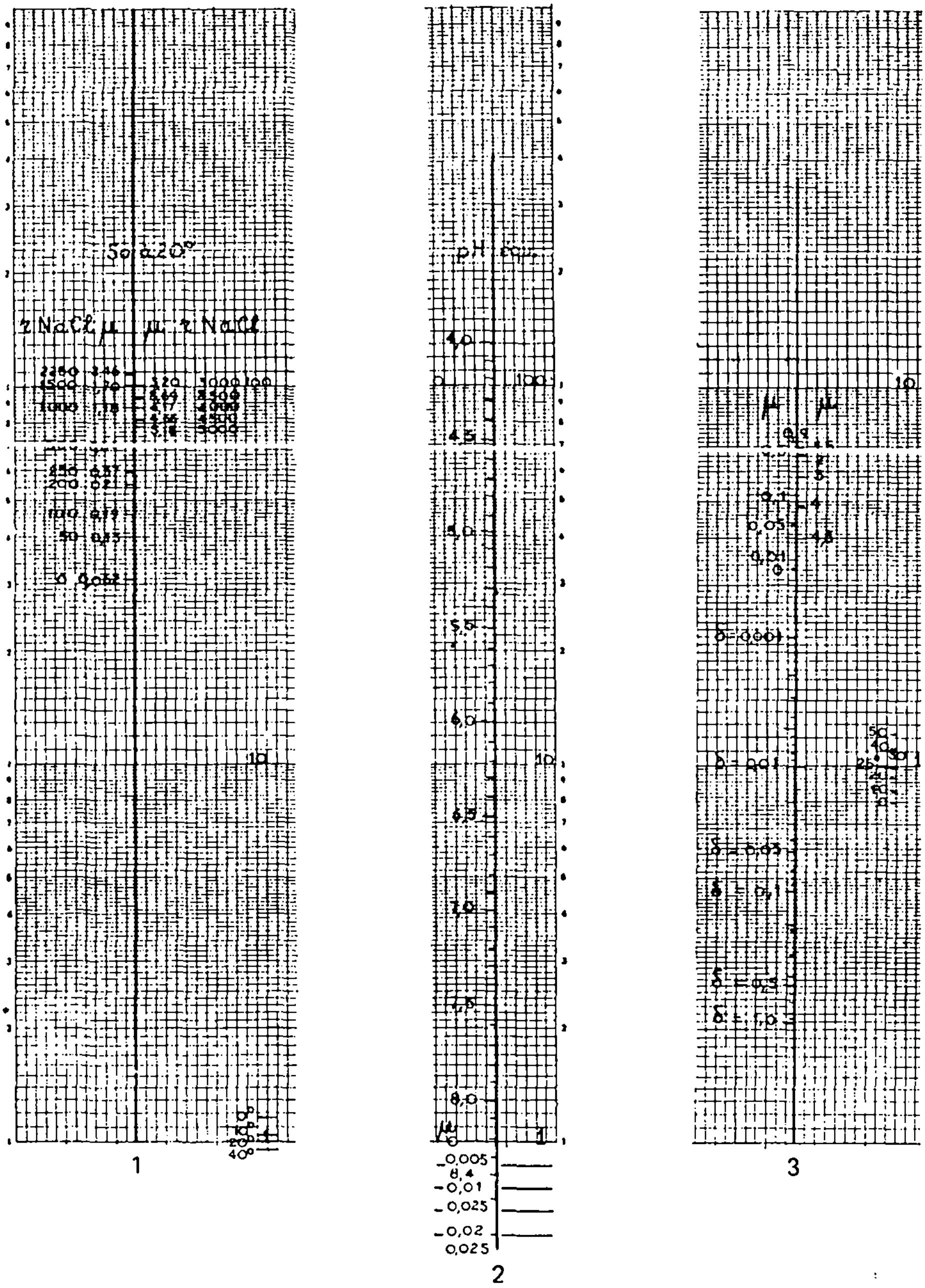


FIG. 7.4.4.4c. Logarithmic scales of solubilities.

1. Scale of solubility of $CaSO_4$, $2H_2O$ for various ionic strengths and various temperatures.
2. Scale of equilibrium pH.
3. Scale of k_r as a function of ionic strength and temperature.

Ionic strength and temperature corrections for pH values

μ	ΔpH	μ	ΔpH	θ	ΔpH	θ	ΔpH
0.001	0.12	0.014	0.44	0°	+0.50	16°	+0.18
0.002	0.16	0.015	0.45	1°	+0.48	17°	+0.16
0.003	0.20	0.016	0.47	2°	+0.46	18°	+0.14
0.004	0.23	0.017	0.48	3°	+0.44	19°	+0.12
0.005	0.26	0.018	0.49	4°	+0.42	20°	+0.10
0.006	0.28	0.019	0.51	5°	+0.40	21°	+0.08
0.007	0.31	0.020	0.52	6°	+0.38	22°	+0.06
0.008	0.33	0.021	0.53	7°	+0.36	23°	+0.04
0.009	0.35	0.022	0.55	8°	+0.34	24°	+0.02
0.010	0.37	0.023	0.56	9°	+0.32	25°	+0.00
0.011	0.39	0.024	0.57	10°	+0.30	26°	−0.02
0.012	0.40	0.025	0.58	11°	+0.28	27°	−0.04
0.013	0.42			12°	+0.26	28°	−0.06
				13°	+0.24	29°	−0.08
				14°	+0.22	30°	−0.10
				15°	+0.20		

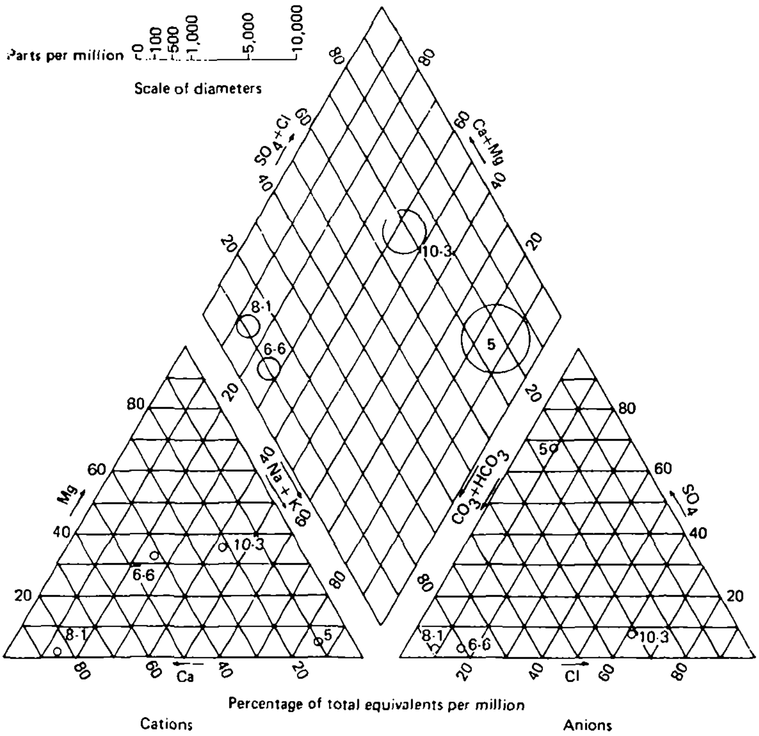


FIG. 7.4.4.4d. Analyses represented by three points plotted in trilinear diagram (after A. M. Piper). (Numbers on plotted points represent sample numbers.)

To obtain the real k_r at an ionic strength μ, the values indicated in the table should be added to the k_r at the point of intersection. Regarding the effect of temperature, allowance should be made for the fact that log k_r decreases by 0.004 for each degree of temperature rise, as indicated on the right of the scale.

If one considers only the major dissolved constituents, and lumps together certain dissolved ions whose properties are similar, most natural waters can be represented as solutions of 3 cationic constituents, calcium, magnesium and the alkali metals (Na and K); and of 3 anionic constituents, sulphate, chloride and those contributing to alkalinity (HCO_3^- and CO_3^{--}). The composition of a water therefore can be represented conveniently by trilinear plotting.

The simplest trilinear plots utilize 2 triangles, 1 for cations and 1 for anions as shown in the lower part of Figure 7.4.4.4d, or a superimposed version illustrated in Figure 7.4.4.4e. Each vertex represents 100 per cent of a particular ion or group of ions. The composition with respect to the cations is indicated by a point plotted in the cation triangle (Fig. 7.4.4.4d) on the basis of the percentages of the three which are present. The composition with respect to anions is indicated by the position of a similar point in the anion triangle.

In recent years a considerable number of publications have appeared in which trilinear plotting of water analyses is described and variations of the basic procedure are suggested. Admittedly the representation of a water analysis by two points each in a separate triangle is not very convenient. Several authors have suggested adding a third plotting field that represents a projection of the triangles into a common area where the analysis can be represented by one point. The position of the single point is dependent upon the

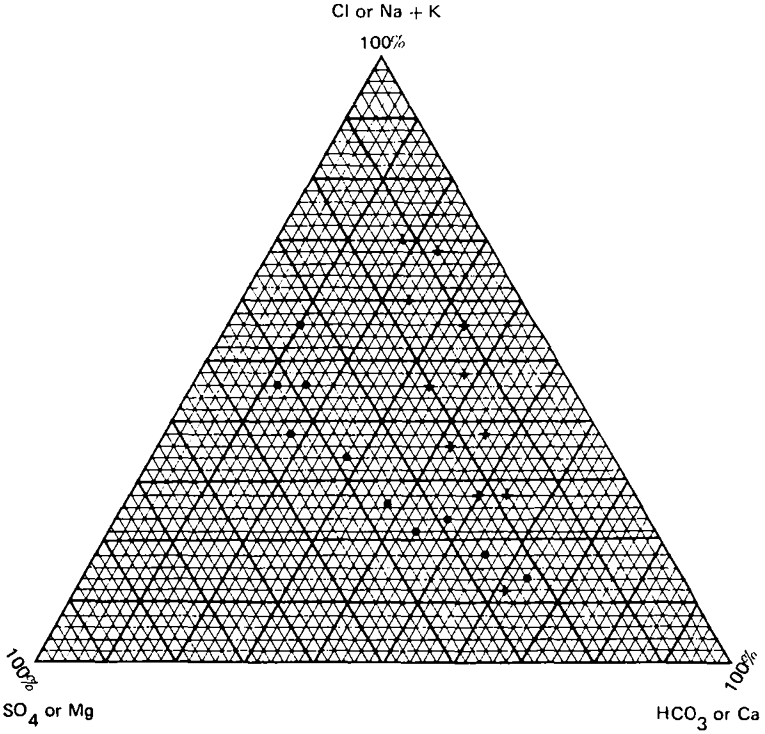


FIG. 7.4.4.4c. Trilinear diagram for comparing chemical characteristics of individual water samples. · Anions; ÷ Cations.

relative concentration of one cation with respect to the sum of the other two, and of one anion with respect to the sum of the other two. Which cation and anion are singled out depends upon the arrangement of the plotting field.

Piper (1944) suggested a form of the trilinear diagram (Fig. 7.4.4.4d) which is extensively used in the United States of America. This diagram represents an analysis by three plotted points. The third point, plotted in the diamond-shaped upper field, indicates the character of the water as represented by the relationships among the Na + K, Ca + Mg, CO_3 + HCO_3, and Cl + SO_4 ions.

Trilinear diagrams (Fig. 7.4.4.4f) constitute a useful tool in water-analysis interpretation. Most of the other graphing procedures represent only a means of pointing up the differences or similarities among waters in a more impressive way than is possible with the numerical data alone. Most of the graphic procedures, therefore, are of value only in pointing out things about the analyses that need closer study, and do not in themselves constitute that closer study which will bring to light the explanation that is needed. By use of trilinear systems the chemical relationships among waters may be brought out in more definite terms than is possible with any other plotting procedure.

Piper (1944) and the other writers who have proposed trilinear graphs have pointed out that, where analyses represent two original waters and a mixture, the analyses will plot on a straight line. For absolute proof of a mixture, the individual concentration values in the analyses must be in proper proportions, and the graphic method alone is not complete proof. However, in many practical problems, this use of trilinear graphs is invaluable. There are many instances in which the possibility of mixture needs to be explored and this procedure is perhaps the simplest one to use in such work. For three-

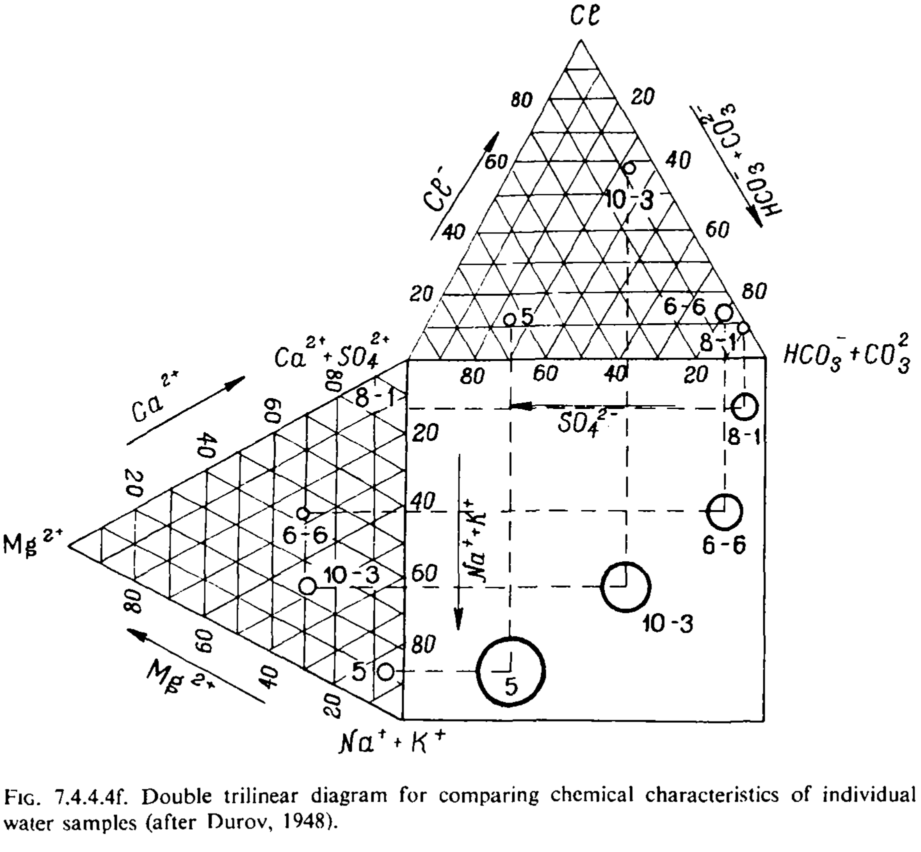


FIG. 7.4.4.4f. Double trilinear diagram for comparing chemical characteristics of individual water samples (after Durov, 1948).

component mixtures, the analysis of the mixture should plot within a figure bounded by the components.

Piper suggests other applications of the diagram. Analyses of waters at different points in a system may plot on a straight line passing through a vertex of the plotting field. If, for example, two analyses line up with the vertex representing 100 per cent Na^+ and 100 per cent Cl one of the analyses could represent water that is the same as the other except for added salt, either from an inflow of water containing a high proportion of sodium chloride or from the solution of solid salt from the rock associated with the water. Orientation towards another vertex could represent, for example, a gain in calcium sulphate, or its loss through precipitation of gypsum.

In certain environments a water may exchange calcium and magnesium for sodium (base exchange) or exchange sulphate for alkalinity (sulphate reduction). Analyses of two waters that are the same except for one of these effects will plot on a straight line parallel to one of the bases of the central plotting field.

For some special studies, it is desirable to show the hypothetical salts composed by the individual ions ($CaSO_4$, $MgSO_4$, $Ca(HCO_3)_2$ and others) in a general plotting field. This can be done by a double trilinear diagram similar to that described by Durov (1948) in which the general plotting field is square rather than diamond-shaped (see Fig. 7.4.4.4f). The Durov diagram is used extensively by Soviet hydrogeologists.

On the whole, trilinear plotting is a very useful interpretative tool. For presentation of data in reports where graphs serve only to illustrate differences among waters, trilinear-plotting procedures are subject to a number of limitations. All such methods emphasize percentage composition rather than total amount of dissolved solids. Concentrations of dissolved matter in individual waters can be represented on the graph by using circles of various sizes instead of points when the analyses are plotted (Fig. 7.4.4.4d) but it is evident that only large differences in concentration will be readily noticeable if this system is used. For those uses where emphasis must be placed upon concentration differences, some other plotting system is desirable as either a supplement or a substitute.

Minor changes in the arrangement of constituents on the trilinear plot may make it more useful for waters of unusual composition. Acid waters and waters that have high enough concentrations of iron, or other unusual cations or anions, to be significant in the over-all composition cannot be shown adequately unless such changes are made. The single-point plotting is dependent only on four variables. In the trilinear diagrams, these are the relative proportions of alkalis ($Na^+ + K^+$), alkaline earths ($Ca^{++} + Mg^{++}$), alkalinity ($HCO_3^- + CO_3^{--}$) and salinity ($SO_4^{--} + Cl^-$). Although these relationships are useful in classifying waters, for some conditions they may not be appropriate. The user may, however, show any grouping desired so long as there are 3 groups of cations and 3 of anions in the lower triangles. The variable represented at 100 per cent at the lower inner vertex of each triangle appears unchanged in the central plotting field. The other two groupings of anions and cations are added to give the four variables represented by a single point in this central field.

For a more complete discussion of the trilinear systems, the reader is referred to the articles by Hill (1940, 1941, 1942), Langelier and Ludwig (1942), Piper (1944) and Durov (1959). A number of examples of applications of the Piper method are to be found in a publication by Piper *et al.* (1953).

In displaying different compositions of water on maps or other illustrations it may be more desirable to use the bar-graph form proposed by Collins (1923) and shown in Figure 7.4.4.4g. Under this system, each analysis is represented by a vertical bar graph whose total height is proportional to the total concentration of anions or cations. The bar is divided into segments to show the concentrations of the cations and anions which make up the total. Usually, there are six subdivisions but more may be used if desired, and each is identified by a distinctive pattern or colour. Cations are plotted on the left

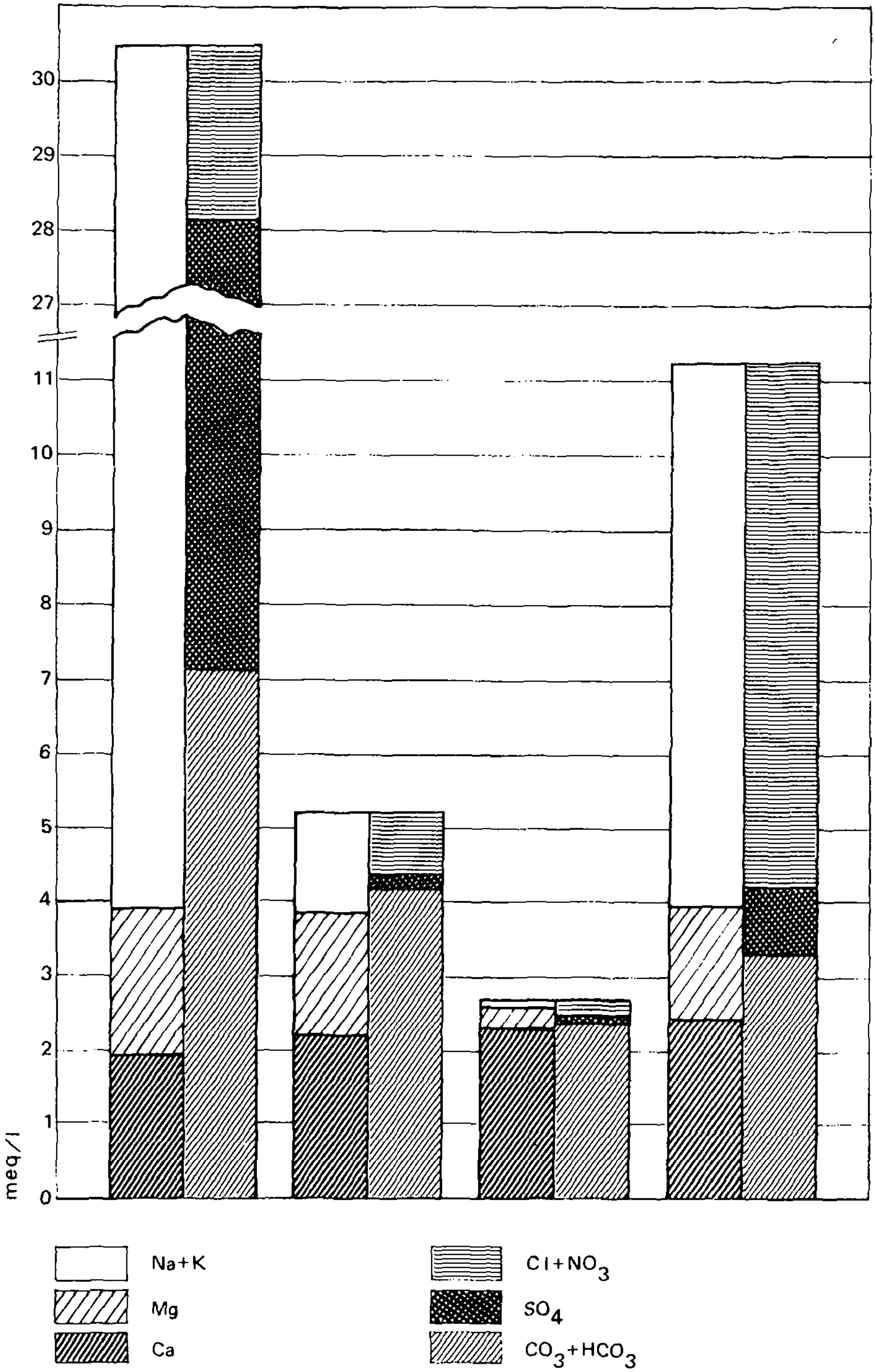


FIG. 7.4.4.4g. Analyses of four water samples represented by vertical bar graphs (after Hem, 1959, p. 169).

half of the vertical bar and anions on the right. Another method of showing chemical composition of ground water diagrammatically was proposed by Stiff (1951).

It should be remembered that the seven illustrations given here (Figs. 7.4.4.4a, 7.4.4.4b, 7.4.4.4c, 7.4.4.4d, 7.4.4.4e, 7.4.4.4f and 7.4.4.4g) presuppose that the principal chemical constituents are those shown. This will be true for a high percentage of the types of water commonly encountered. The investigator should recognize, however, that there are special environments, either natural or artificially induced by man, that can produce water types in which F, NO^3, PO^4 or Fe, or some combination thereof, represent a major portion of the anions. In these circumstances the diagrams or graphs would have to be labelled accordingly.

When detailed investigations of the chemical composition of ground water are made, hydrochemical profiles and graphs are compiled showing the change in the concentration of the principal dissolved constituents in space and time. Besides the chemical composition of waters, data on water levels, ground-water discharge and hydrometeorological factors are also shown in graph form. Changes in the chemical composition of ground waters with time are shown by curves constructed for selected individual cations and anions and for the total dissolved-salt content (commonly plotted as milli-equivalents per litre along the *y*-co-ordinate axis). If all the graphs are constructed to the same time scale they should reveal changes in concentration of individual constituents both in time and space and also changes in the ratios of the various anions and cations.

7.4.5 *Observation of soil moisture in the zone of aeration*

Observations of soil moisture in the zone of aeration are made *in situ*: (a) to calculate the moisture content and its change with time, in the soil zone between the land surface and the water table; (b) to compute the soil-moisture balance in this zone and to study its relationship with the ground-water balance; (c) to follow moisture movement in the zone of partial saturation under the influence of different factors; and (d) to study moisture dynamics and the transport of water-soluble salts.

The choice of soil-moisture observation stations must be made with care and should be guided by the following principles:

1. Stations should be located where the geomorphological and general ground-water conditions are typical of the regions selected for detailed water-balance study. The sites selected will in turn reflect the effect of such geomorphological features as alluvial terraces of rivers; long gentle slopes of river banks; relatively flat watersheds with various depths to the water table; and undulating terrain or unevenly ridged relief.
2. Station locations should favour water-balance plots where observations of other elements in the water balance at the land surface and for the ground-water régime are being or will be made.
3. When there are no such plots or balance sites in a basin, and only observation wells with water-level recorders exist, moisture-observation stations are established in accordance with the computed elements of water flow.
4. Depending on the method of soil-moisture observation, the study plot size can range from a few to several-hundred square metres.
5. Station locations reflect the objective of the particular experiment but they should also take into consideration the manner in which soil moisture is affected by the character of vegetation, the nature of the soil, the microrelief, the agronomic methods being used in the cultivated fields and the changes in ground-water factors with time and depth. For example, establishment of soil-moisture stations in forest regions should include locations at the forest edge, inside the forest, in clearings and in open fields outside the forest. This enables comparison between various moisture régimes and water balances under a broad range of conditions.

Two methods are commonly used for soil-moisture determination, namely, weighing and drying, and neutron-scattering techniques using one of several suitable radioactive isotopes. The first method, though more time consuming and inconvenient for small plots, permits study of the problems referred to above but unfortunately the site is destroyed in the process. The second method is convenient to apply on small plots of homogeneous soil structure and enables repeated measurements in the same hole of the soil-moisture régime at different depths.

In applying either method the frequency of observations to be maintained should be established beforehand by a preliminary sampling programme over a specific period of time, and making a statistical analysis of soil-moisture observation results in the sampled area of the basin.

The frequency of soil-moisture observations is determined primarily by the objectives of the programme. For investigating the occurrence of shallow-lying ground-water systems, important to general water-supply development, or for the solving of problems concerning land reclamation (irrigation, drainage), observations during the first few years can be made monthly: e.g. before and after heavy precipitation or irrigation. Observations taken four times a year may be adequate for seasonal water-balance computations. For long-term investigations of the ground-water régime and water balance under natural conditions, two observations a year may suffice, e.g. at the end of cold and warm seasons. However, it is desirable to have data on soil-moisture content before and after significant rises of the water table in spring or autumn.

It is important to establish the permissible error and confidence interval for soil-moisture data. For this purpose the required number (n) of observations on soil moisture at a particular depth depends upon a number of factors including the natural variations in soil-moisture data, the stipulated permissible error, and the required confidence level. An accuracy of 10 per cent in soil-moisture determination and a confidence level of 90 per cent will generally suffice.

A series of bulk soil samples is taken over a one- or two-day period close to the selected observation station in order to determine values of the soil-moisture content and the volume-weight relationships of the soil. To obtain a sufficiently large number of samples and to determine the arithmetical mean moisture content with a high degree of accuracy (error less than 10 per cent) will probably require drilling about 30 wells, this number being required for each genetic and lithologic variety of soil. Subsequent additional observations will be made at the depth from which the samples are taken.

The percentages of water content by weight, V_{wt}, obtained by weighing and drying the bulk soil samples, may be converted into percentages of water content by volume, V_{vol}, using the relation

$$V_{vol} = \delta \, V_{wt} \qquad 7.4\,(2)$$

where δ is the density of the dry soil. These percentages, for each series or set of soil samples, are analysed statistically by the following procedure:

Compute x, the arithmetical mean, d, the absolute deviation from that mean, and S, the standard deviation, from the relation

$$S = \sqrt{\frac{\Sigma d^2}{n-1}} \qquad 7.4(3)$$

in which n is the number of samples.

Determine $S_{\bar{x}}$, the standard error of the mean, from the relation

$$S_{\bar{x}} = \frac{S}{\sqrt{n}} \qquad 7.4(4)$$

Determine C, the coefficient of variation (as a percentage), from the expression

$$C = \frac{S \times 100}{\bar{x}} \qquad 7.4(5)$$

and p, the accuracy of the test, from the relation

$$p = \frac{C}{\sqrt{n}} = \frac{S_{\bar{x}}}{x} \times 100 \qquad 7.4(6)$$

The foregoing procedure can be illustrated, for a set of observations, by assuming that the fiducial limits of the mean shall be no greater than 10 per cent. In other words, the deviation from the true mean (m) should be no more than ± 10 per cent of that mean. The pertinent parameters are related as shown by the equation

$$\frac{0.1\, m}{\sigma} = \frac{t}{\sqrt{n}} \qquad 7.4(7)$$

where σ is the true standard deviation and t is called the standard unit of the deviation expressed in units of $\sigma_{\bar{x}}$. In this example the true mean and the true standard deviation are not known; thus $\bar{x}$ must be used for m and S for σ. With these substitutions a value may be computed for the left side of Equation 7.4 (7). From column 6 of Appendix G find a value equal to or less than the preceding computed value; then move horizontally across to column 1 to find the number of observations required to stay within the specified fiducial limits.

When no bulk sampling is done to determine moisture content in the upper 1-m soil layer, 4 or 5 samples are taken in the upper layers and 3 in deeper strata.

To determine soil-moisture content, samples can be taken at a depth of 5 cm and successively deeper at 10, 20, 30, etc., cm down to a maximum depth of 1 to 2 m; below this, down to the top of the capillary fringe, samples are taken every 20 to 25 cm. Within the capillary fringe samples are again taken at intervals of 10 cm. This sampling pattern, to determine soil-moisture content, provides good accuracy for calculating moisture and saturation deficits.

Volume weights of soil are commonly determined only once (with the number of samples depending on the soil-moisture content) for every characteristic layer or depth interval during the first series of measurements. Core samplers or special steel rings are used to determine volume weight. These are driven into the soil for each layer to be measured.

On small plots (of the order of 4 to 6 m^2) the boreholes are spaced in such a way as to ensure uniform distribution of observations over the total area for any given set of readings. Distances between the holes drilled at different times may range from 50 to 100 cm. After sampling, a hole is carefully plugged layer by layer with the extracted soil, and the boring site is marked.

Schematic examples of hole arrangement with four, five, and sixfold sampling and four, five, and six sets of observations per year are given in Figure 7.4.5a. The numbers in the squares represent the sequence of soil-moisture determinations.

1	2	3	4
3	4	1	2
1	2	3	4
3	4	1	2

I

1	2	3	4	5
4	5	1	2	3
2	3	4	5	1
5	1	2	3	4
3	4	5	1	2

II

1	2	3	4	5	6
4	5	6	1	2	3
6	1	2	3	4	5
3	4	5	6	1	2
5	6	1	2	3	4
2	3	4	5	6	1

III

FIG. 7.4.5a. Arrangement of boreholes on microplots for soil-moisture sampling.

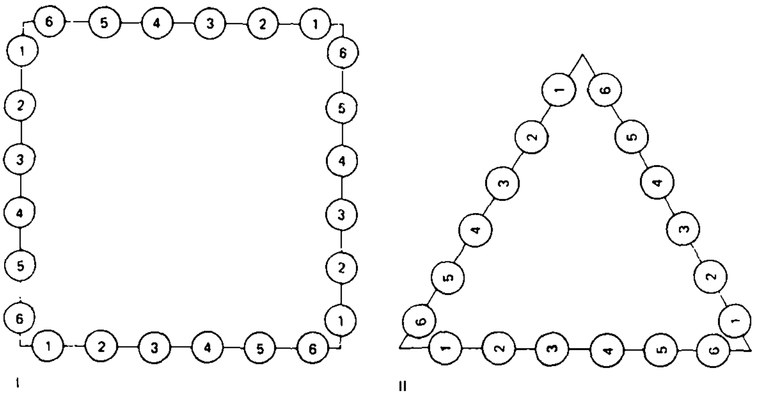


FIG. 7.4.5b. Location of boreholes for sampling along perimeters of polygons. I, Location of boreholes for six sets of observations with four samples per set. II, Location of boreholes on microplots for six sets of observations with three samples per set.

When soil-moisture observations are made in agricultural fields where the borehole locations must be restricted to the field perimeters the successive sets of observations are made in the sequence shown by the closed polygons in Figure 7.4.5b. The distances between holes are 1 to 2 m and the maximum length of a polygon side is thus 2 *n*, where *n* is the number of sets of soil-moisture determinations.

Sampling for soil-moisture determinations is generally done with a hand auger; normally one or two men can drill holes to a depth of 3 to 5 m. The bits of such augers are chosen with respect to the soil structure and compactness.

Use of radioactive isotopes in the neutron-scattering technique involves a neutron moisture meter. Americium and beryllium are the most common neutron sources. The meter consists of (a) a pulse amplifier; (b) a probe (radioactive-source holder); (c) a shield and calibrating or standardizing device; and (d) a counting device. This subject is discussed in two publications by IAEA (1968, 1969).

Both publications offer more detailed discussions than the summary in the following paragraphs.

The soil-moisture meter operates on the principle that fast neutrons emitted from the neutron source are involved in repeated elastic collisions with nuclei of the elements in the surrounding materials. The neutrons bounce off heavy nuclei with most of their original energy intact; but collision with lighter nuclei transfers a greater fraction of the neutron's kinetic energy. Thus substances containing hydrogen (such as water or oil), which has the lowest atomic weight among the elements, are the most effective in causing the fast neutrons to lose energy rapidly and become thermal neutrons. The flux of thermal neutrons returning to the probe by diffusion processes is measured by the counter. This flux is proportional to the amount of hydrogen incorporated in the materials surrounding the probe and thus—for water-bearing materials—is related to the moisture content. The probe response depends upon the volume of surrounding material sampled; normally a spherical shape about 35 cm in diameter, but with low moisture content in low-density materials this diameter may be as large as 120 cm.

Calibration curves have been drawn so that by entering the ratio of pulses counted in the soil to pulses counted in the standardizing shield, soil-moisture content by volume can be determined, in certain circumstances to an accuracy of ± 2 per cent.

In addition to the normal meteorological observations made at a water-balance site, the following records should be collected along with the soil-moisture records:

1. Temperatures in the zone of aeration from the land surface down to the water table, with measurements at depths of 5, 15, 40, 80, 160, 320 cm and deeper as appropriate.
2. Evaporation from the surface of soil and water using weighing evaporimeters.
3. Temperature and humidity of the lower atmosphere at heights of 50, 100 and 200 cm above the ground.
4. Position of the water table in observation wells, for which a continuous record of the water-table fluctuations may be appropriate.

7.4.6 *Observation-well management*

To achieve valid observations of the ground-water system it is necessary to exercise systematic inspection and control over the observation network and the work of the technicians making the observations.

In order to be sure the data obtained are representative of the point being sampled, observation wells, pits and springs should be cleaned and pumped periodically, generally once a year. After cleaning the water pumped should be clear and the screen section clean. Wells and pits should be sounded periodically to check for damage by clogging. If clogging is discovered, immediate measures must be taken to clean the well and repair the damage.

The sensitivity of a well can be checked by pumping or recharging water into the well. Observations are then made of the rate at which the disturbed water level returns to static. The actual rate of water-level rise of fall, compared with the aquifer's theoretical response, will indicate whether or not the well's intake area is in need of cleaning.

While inspecting the observation points the condition of the measuring equipment and instruments and the accuracy of datum corrections of the measurements should be checked. This is especially important when using wells that are in active use for observations of regional water level because the measuring point elevation may be altered by maintenance on the well.

When using automatic water-level recorders their operation is checked by measuring the water level manually and comparing the result with the recorded value. The measured depths to the water are noted at the beginning and the end of the recorder record and adjustments can be made to correct the record for instrumental error in both the vertical and time sense.

7.5 Records and data display

Complete records of the basic data collected by any hydrogeological organization must be kept at some central location and procedures established to ensure the accuracy of any information added to the permanent records. Much long-term benefit can be realized if the data collection system is organized so that digital computer techniques can be used to handle and manipulate the data being obtained.

All observation posts must be levelled before meaningful comparisons of the data can be made and the heights of the measuring points (i.e. the points from which measurements are made) should be systematically checked. In this connexion, it is desirable to set bench-marks near every separate observation point or group of points so that measuring point elevations can be readily recovered.

Periodic relevelling of the measuring-point elevations should be done if there is any reason to suspect that the casing is not firmly set in the borehole. New wells may sink slightly until the casing establishes a firm bond with the borehole walls or else may rise slightly, under some conditions, as a result of frost heaving.

The water-level measurements made by the technician visiting the well must be recorded in a field notebook at the time they are made. The data on all wells obtained on a certain date are put in succession into the field notebook, including the time at which the measurement was made. The office records are updated by transferring the field data to the permanent records at convenient time intervals. Periodic water-level observations are recorded on office data sheets which are arranged in time sequence by individual well. If the well flows, observations of yield are generally included with the head measurements.

On the basis of the periodic observations yearly lists of observation results are compiled. In these tabulations data on depths to water or absolute marks of water level are given. Much of this compilation work is now done by digital computer.

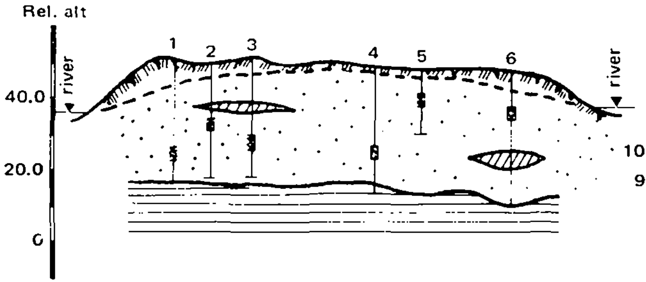
The office record of a well should also include data such as: well log, hydrogeological conditions (the depth to the aquifers encountered in drilling and their respective water levels), specific discharge, chemical composition of water, period of time between measurements, design, depth, absolute elevation of the measuring point, etc.

The results of observations of infiltration and evaporation for all lysimeters are entered in a field notebook in successive and chronological order. Calculations of infiltration and evaporation values are made and recorded in office records. For these calculations data on the specific yield and saturation deficit are obtained.

When observing soil moisture in the zone of aeration by means of a neutron moisture meter, a standard measurement of the number of impulses is carried out to provide calibration data and then impulses are counted through the total thickness of the zone of aeration at depth intervals chosen in advance. An average number of impulses is calculated on the basis of three readings taken at every point. Soil moisture is determined with the help of a calibration curve relating the average number of impulses at every depth of measurement to the equivalent moisture content.

The investigation of any area requires the systematic organization and presentation of the data being collected. This is best done by the use of maps and graphs. The following maps are among those useful in any investigation:

1. An areal map showing the location of observation posts, (ranges and single observation points), and meteorological stations and gauges. The map should be systematically updated.
2. A geological map on a convenient scale.
3. A geomorphological map on the same scale.
4. A hydrogeological map on the same scale.



▨ Soil layer of loamy sand with plant roots and gravel.

⁘ Sands, yellow-brown, greyish-yellow and grey, fine to coarse with gravel and pebbles.

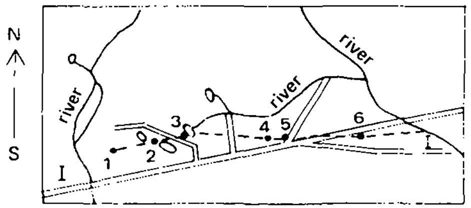
▨ Loams, yellowish-brown and yelloish-grey, loose.

☰ Mesozoic deposits: clays, black and dark-grey, solid, slightly micaceous, sometimes with concretions of phosporites and fauna.

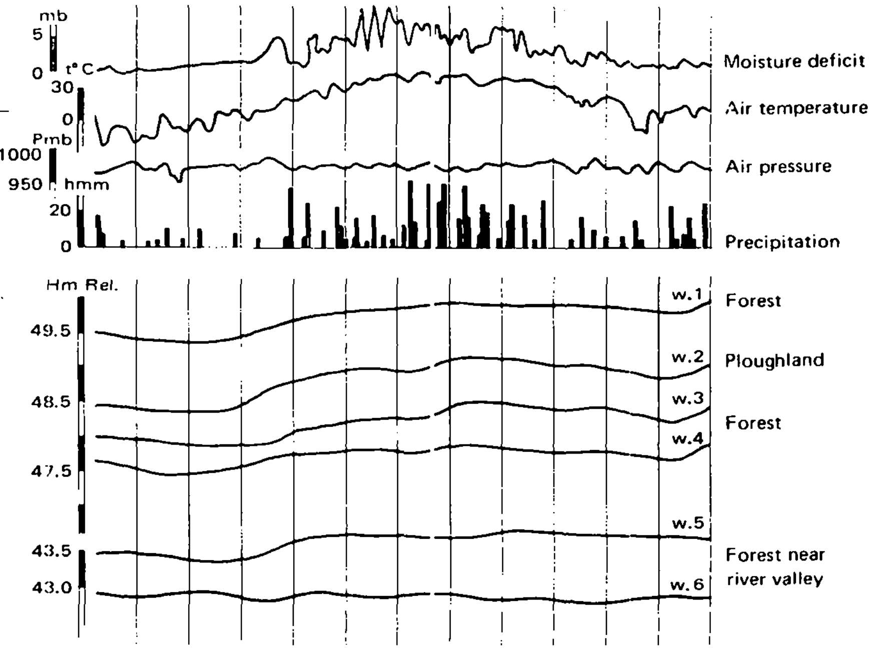
– – – – Water table.

▮ *Depth of screen installation.*

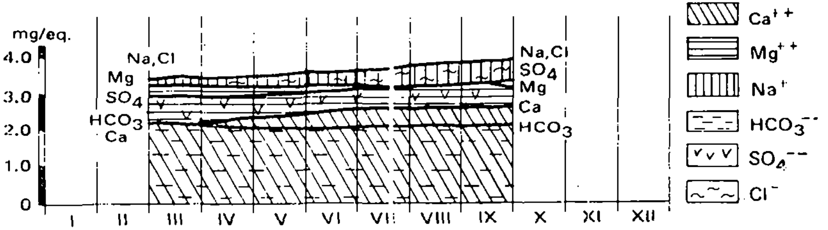
Hydrogeological section



Location of observation wells



Ground-water level fluctuations in 1965



Chemical composition graph for well 2

FIG. 7.5a. Graphs of ground-water régime.

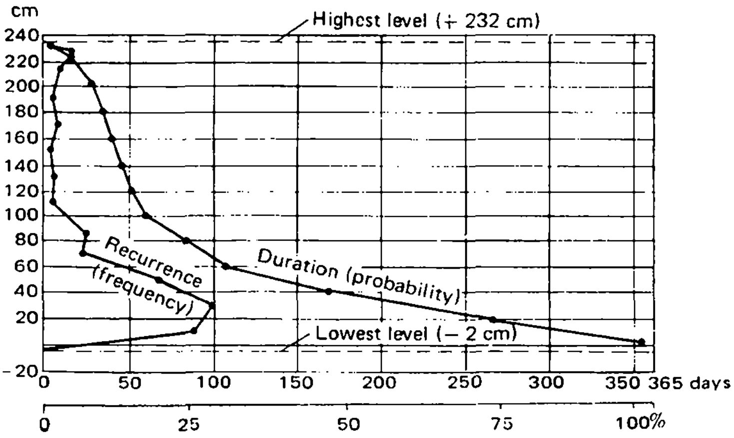


FIG. 7.5b. Frequency and duration of ground-water levels.

5. A map of factual material reflecting, besides observation ranges and single points, those points where experimental and exploratory work was carried out. The scale of such a map is determined by the amount of data to be plotted.
6. A map of the total dissolved solids content and chemical types of ground waters compiled from available data. As data are accumulated such maps can be compiled for different years and seasons, presuming seasonal fluctuations are found to occur.
7. A map of the depths to ground water and ground-water contours compiled according to the existing data. For areas where detailed observations are available, large-scale map fragments can be used to illustrate this detail in the areal picture.

The following graphic materials should also be available as aids in interpretative ground-water studies:

1. Stratigraphic columns of all observation and test wells.
2. Graphs of observations of the water level, discharge, temperature and chemical composition of ground water (Fig. 7.5a). It is desirable to time-relate these graphs with the chronological graphs of the régime-forming factors, i.e. graphs of temperature and air-humidity deficit, atmospheric pressure, precipitation.

Such graphs should be drawn up for a whole range of observation wells. It is convenient to draw water-level fluctuation graphs on comparable scales to facilitate comparison. Figure 7.5a is an example of such graphs.

While processing observation-well water-level data the frequency and duration of water levels can be calculated and the corresponding graphs are constructed (Fig. 7.5b).

Forms of documentation designed to facilitate exchange of information on observation results were adopted at the second session of the IHD Co-Ordinating Council held in Paris from 19 to 25 April 1966. They may be found in the appendixes to the report of this session.

7.6 Planning and design of ground-water level networks[1]

7.6.1 *Aims of network design*

The purpose of ground-water network design is to allocate available funds, manpower, equipment and time so as to obtain in the most efficient manner possible the ground-water information needed for all aspects of water-resources programmes. By means of a well-designed network, the data needed for varying requirements can be supplied readily at minimum cost.

In order to design networks for data collection, it is first necessary to identify the various components of the hydrological system and to determine their relationships. The next step is to decide how to collect the data in the most economical and satisfactory manner.

The scope of the system must be broad enough to provide for international, regional, national and local needs. Local needs usually require more detailed coverage than do the national interests.

7.6.2 *Definitions of terms*

Network : A systematic pattern of observations.

Water-data network : A network that provides information for the planning, development, management and administration of water resources and water-related activities.

Ground-water data network : A water-data network giving special importance to the collection of ground-water information.

Ground-water-level network : A record of systematic measurements of ground-water-level changes over a period of time.

7.6.3 *Data collection*

In the broad context, the networks require the collection of data to establish (a) fixed parameters, with a strong emphasis on area sampling; and (b) varying parameters requiring sampling over differing time spans, which can be used as inputs for the development, testing and verification of models. One objective is the development of a model which could be used to predict the trends in ground-water behaviour with respect to both quantity and quality, and in response to natural and man-made stresses. This could then serve as a basis for future planning and operational needs. The objective may never be attained, but approximations are very useful.

In order to meet a variety of needs, a network should consist of two parts, involving the collection of information (a) that satisfies immediate and specific needs, and (b) that will meet over-all requirements or future specialized needs as they arise. The first subdivision entails rational planning of the network to meet needs that are already known. The second involves long-range planning, design and implementation, based on both theory and engineering experience. This subdivision may exist at several different scales, thus requiring the collection of information at the international, regional, national and local levels. This approach is especially needed in large nations with diverse terrains,

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[1] *This section has been prepared by* G. H. Davis R. C. Heath, P. R. Seaber and others of the United States Geological Survey and revised by a panel of editors composed of: A. A. Konoplyantsev, E. P. O'Driscoll, H. J. Schoeller and G. C. Taylor.

differing population densities, varying levels of economic development and consequent political and social demands.

National network design is more than a choice of specific observation points. Design cannot be expressed completely as points on a map, because ground-water information relating to depth, extent, quality, quantity, movement, availability, use and time variability is also included. The amount of information needed in any area is dependent on the present stage of development and the problems involved, as well as variations in the geohydrological environment. The amount and type of information needed can be determined from field experience, information already available and obvious needs for new facts.

Field data should be collected for three purposes: accounting, surveillance and areal synthesis of the system. Accounting is a process whereby a progressive record of one or more components of the system is kept as, for example, the amount of water stored in an aquifer throughout an interval of time. Surveillance entails monitoring a component to ensure that it does not vary in quantity or quality from the required standard, or create a hazard. Areal synthesis includes the separation and study of spatial distribution or behaviour of a component of the system, in order to gain a better understanding of how it is related to other components and to the environment. These three processes, although distinct, are interrelated and mutually supporting.

Different levels of development may need different types of information, even though these are all part of the hydrological cycle. A planning programme or network design should identify those elements of ground-water hydrology which are basic to a thorough understanding of the system, and then should design the field programmes of data collection so that these elements can be studied. Most data are collected for a particular purpose, which usually involves a local problem. However, programmes must be designed so as to present the regional, national or international view of a ground-water system, while using the data and information provided by local studies. Each programme should provide added experience, useful in future planning.

7.6.4 *Ground-water data networks*

Activity networks. Networks may be classified according to the types of field data collected, including water-level measurements, water use, water quality, pumpage and geological data. They may include the whole range of geohydrological information necessary to describe the natural system in space and time.

Networks for special programmes. Networks may be classified according to their objectives, in either of two ways: (a) by economic, social, developmental and environmental needs, or (b) by category, or functional demands.

Economic, social, developmental and environmental needs. Networks may be useful in assessing the relative importance of economic efficiency, social well-being, regional development and environmental quality.

Many countries are progressing beyond the stage of developing water resources on a local or river-basin scale, and are planning broad regional or nation-wide projects. These projects require the collection of basic data at a national level. Beyond this there remain the traditional differences and needs for data which may be required in more detail or with more urgency by development projects in different parts of the country. To allow for these differences and to make the network design more flexible, three principal stages in development may be identified. These are: (a) exploration; (b) exploitation; and (c) conservation. Different parts of a region or nation may be in any of these stages at any given time. Technological, political, sociological and economic changes may also bring about rapid shifts from one stage to another. The three stages mark the shift in degree of usage from: (a) an essentially untapped resource; through (b) an active development stage; to (c) a third stage when problems created by development must be understood and controlled. The third stage occurs when any stress placed on the system necessitates the ability to predict the resulting strain.

Economic, social, developmental and environmental needs for particular uses of water and related land-resources often require special networks. These include, for example, networks designed to answer questions related to water-quality control, recreation, water supply, flood control, land treatment, geothermal power, navigation, drainage, mining, reservoirs, canals, mineral waters, etc.

Category or functional demands. The rationale for the design of a ground-water data network rests upon the identification of water-data demands and the supply of information needed to meet these demands. The gap between supply and demand for information should be identified, and a data programme formulated to bridge the gap. Information is needed mainly for planning, implementation, management, research and administration. One fundamental step in network design is to decide what are the levels of information needed by the potential users. The demands evolve in accordance with the general administrative structures and with their interests and concerns at different governmental and private-sector levels. This involves determining the quantity and quality of information needed to answer questions that arise at the international, national, regional, local or private level. The network must be specifically designed to answer questions that arise during development or from the areas of interest and financing.

The first or nation-wide base level of information is used mainly for broad national planning and assessment, to lessen the impact of unanticipated needs, and to provide a base for more detailed and more precise future work. The information obtained should be sufficient for a gross estimate of the ground-water resources at any place at any given time, if suitably processed. This level should be set high enough to provide data for an areal synthesis in general terms of the hydrological system, for the periodic accounting of water use in the major national regions, and for the surveillance of water-related hazards. Data acquired at this level would be used mainly for periodic assessment of resources and environment.

The second level of information meets the needs of water-resources planning and assessment for a whole region, such as a large relatively uniform ground-water province. This level would add to the first or basic level those data that are specific to the needs of each region, taking into account the urgency of needs, and the accuracy and detail which is required. This level should be adjusted for each region in consultation with local authorities.

The third level of information is comprised of data mainly utilized for local water-resources operation and management. As water use intensifies, the need for this level of information increases, so that data for specific operational, legal or administrative purposes may be provided. Information is needed urgently for local metering of water quantity or quality.

For the first two levels of information, the network should be designed to anticipate specific needs. In contrast, a local network need not be part of a specific national programme, but should meet local needs and involve local decisions. The several types of networks fulfil different requirements and require various levels of data; they cannot all have the same design.

The following sections focus on the design of programmes for observations of ground-water levels at the regional and local levels. Other types of ground-water data are collected to provide information on aquifer inhomogeneity (Altovsky *et al.*, 1959), and time-variable properties, such as water use, temperature and chemical quality, and these are treated in other chapters.

TABLE 7.6.5. Objectives of observation-well programmes and the frequency of data collection to meet the objectives

Objectives of observation-well programmes	Frequency of data collection to meet objectives											
	Periods without fluctuation	Earthquakes	Atmospheric pressure	Earth and ocean tides	Man-controlled loading	Cyclical withdrawal or injection	Precipitation	Stream and lake fluctuation	Seasonal changes in climate	Long-term changes in climate	Progressive changes in withdrawal	Unanticipated changes in withdrawals
Effects of stresses on recharge to and discharge from ground-water systems	1					3	2, 3	3	2, 3	2	2	4
Status of ground-water storage	1								2	2	2	4
Hydraulic characteristics of ground-water systems: transmissivity	1			3		3		3				
Storage coefficient						3	3	3				
Degree of confinement in a ground-water system		3	3	3	3	3	2, 3	3			2	4
Areal extent of aquifers in a ground-water system	1					3					2	4
Research						Depends on objectives of research						

1, single measurement at many points repeated at intervals of months or years; 2, measurement at daily, weekly or monthly intervals for long periods; 3, continous measurements for short to medium periods; 4, continuous measurements for long periods.

Source: R. C. Heath, 'Why Measure Ground-water Levels?' *U.S. Geol. Survey Water Resources Bulletin*, January-February 1974, p. 457-72.

7.6.5 *Ground-water-level networks*

The principal objectives for systematic measurement of ground-water levels and the frequency of collection of data (Konoplyantsev *et al.*, 1964) are summarized in Table 7.6.5 and are discussed in more detail in the following sections (Heath, 1974).

7.6.5.1 **Effects of stresses on recharge to and discharge from ground-water systems**

Theis (1937) has pointed out that unless and until withdrawals from a ground-water system are balanced by an increase in recharge or a reduction in natural discharge, water will be removed from storage and water levels will decline. Therefore, the rate at which water can be removed from a ground-water system for an indefinitely long period (i.e. what some hydrologists would call the perennial yield) depends on the extent to which the withdrawals affect recharge and discharge conditions. At present, there are relatively few systems for which the effects of withdrawals can be predicted reliably, and it is necessary to watch how they behave.

Prediction of perennial yields depends on a knowledge of how the drawdown behaves in the two different pumping situations of (a) constant heavy withdrawal; and (b) gradual or intermittent increases in the pumping rate.

A correction must be made for water-level changes not caused by pumping but by other factors, such as seasonal and longer-term variations in climate. These climatic changes constitute natural stresses on ground-water systems, which may provide useful clues as to how the systems will respond to the more pronounced stresses of large withdrawals. In fact, where a ground-water system has not yet been heavily pumped, the stresses resulting from variations in climate may provide the only measure of the system's potential behaviour under pronounced stress.

Predicting how stress will affect the ground-water system is by far the most important objective of a water-level network.

7.6.5.2 **Status of ground-water storage**

In many arid and semi-arid regions, a large part of the water withdrawn from ground-water systems comes from storage. A periodic review of the storage and its rate of depletion are most important for water management. The principal symptom of depletion is a progressive decline in water levels; with the exception of local areas of serious overdraft, this is usually not observed in ground-water systems in regions of high rainfall.

7.6.5.3 **Hydraulic characteristics of ground-water systems**

The most important hydraulic characteristics of a ground-water system are its capacity to transmit water (its transmissivity) and its capacity to store water or to release water from storage (its storage coefficient). Knowledge of transmissivity is essential for nearly all predictions of well and aquifer yields, and is one of the parameters that must be included in digital and analogue models. The storage coefficient affects the rate of response of a ground-water system to stress. Knowledge of the storage coefficient is essential for the prediction of drawdowns under transient conditions.

Both transmissivity and the storage coefficient are usually determined by test pumping of the aquifer, but may also be determined from water-level fluctuations in a network of observation wells. These are indicated in Table 7.6.5, and they include the analysis of water-level maps prepared from a quasi-simultaneous series of measurements made in observation wells throughout a given area or region, and an analysis of water-level fluctuations caused by tides and cyclical pumping operations.

7.6.5.4 **Degree of confinement in a ground-water system**

The response of an aquifer to stress depends largely on whether the aquifer is confined or unconfined; if confined, the hydraulic conductivity of the confining beds also plays a role. The degree of confinement is indicated quantitatively by the storage coefficient, and qualitatively by the response of water levels to short period phenomena such as earthquakes, changes in barometric pressure, tides and the passage of railway trains. It may also be indicated by water-level fluctuations resulting from individual rainfalls, and by fluctuations caused by changes in stream and lake levels.

An indication of the degree of confinement is the minimum information that should be obtained from water-level measurements in an observation well. In order to obtain such information it may be necessary to operate a graphic recorder on the well for a short period.

7.6.5.5 **Extent of areas underlain by aquifers in a ground-water system**

The yield of a ground-water system depends on the area and the hydraulic continuity of its aquifers, as well as the hydrological nature of their boundaries. The extent of some aquifers, for example alluvial deposits or glacial outwash, may be readily apparent from surface observations or from geological and topographical maps. However, the extent of other aquifers, such as those underlying coastal plains or interior lowlands may not be obvious even from detailed stratigraphical maps. The most practical means of determining the areal extent of such aquifers is to observe the continuity of the ground-water levels. Classic examples are the water-table map of Long Island, the maps of the potentiometric surfaces of the regional limestone aquifer of Florida, and the Roswell artesian basin in the United States of America. Many other examples from other parts of the world might also be cited. Water-level data showing regional cones of depression around major pumping centres provide similar information.

Determination of the regional continuity of water levels requires water-level measurements made over a relatively short period of time in a large number of wells (so-called 'mass measurements') or, when this is not feasible, the adjustment to a comparable base of water-level measurements made at different times. The extent of aquifers (or what might more properly be called their 'uninterrupted hydraulic continuity') may also be determined from water levels measured at widely spaced observation wells, as demonstrated by Legette (1937) on Long Island, when he observed the drawdown and recovery of water levels in an artesian observation well more than 10 km from a pumping well.

7.6.5.6 **Research**

Water-level data from observation wells are also used in research projects not directly related to determining the yield of a ground-water system. These projects include, among others, studies of land-surface subsidence, the prediction and control of earthquakes, and the effect of channels and other man-made projects on the ground-water system.

7.6.6 *Water-level observation well programmes*

Several objectives for systematic collection of ground-water levels can be met by the use of the same well or group of wells. The design of an observation well programme should meet these objectives by providing three distinct networks: (a) a hydrological network showing regional ground-water levels; (b) the base-line network, indicating the response of water levels to climatic and other natural changes; and (c) a water-management network, demonstrating how the ground-water systems react to artificial stresses.

7.6.6.1 Hydrological network

The preparation of a regional water-table or potentiometric-surface map is the first step in understanding how a ground-water system functions. Such maps show the direction of ground-water movement and help in identifying the extent of aquifers and the recharge and discharge areas. They are also invaluable in predicting the response of aquifers to stresses and in analysis of changes in ground-water storage. If the observation points are sufficiently close, they may also show the extent of cones of depression.

A hydrological network consists of a group of observation wells, more or less randomly located over an area that may range in size from a few tens to a few thousand square kilometres. The position of the water level, relative to sea level or some other datum, is measured (simultaneously) in all the wells. The interval between successive measurements may be as short as a month in a system in which the water level changes rapidly, to a decade or more for large systems not yet intensively developed.

The number of observation wells needed per 1,000 km^2 depends on the geohydrological complexity and the amount of detail required. Well densities may range from 100 per 1,000 km^2 for a complex area for which considerable detail is desired, to as few as 1 or 2 per 1,000 km^2 for a large area in which mapping of only major features is required. All wells recording one type of parameter must be open to the same water-bearing zone.

7.6.6.2 Base-line network

A base-line network consists of observation wells situated in an area where water levels are not significantly affected by withdrawals or other major artificial stresses. Data from this network show how the water levels respond to variations in climate and other natural phenomena, and thus provide basic information for the interpretation of data from both the hydrological and water-management networks. Some of the factors which must be considered in a base-line network are outlined in Table 7.6.6.2 (Konoplyantsev *et al.*, 1963).

Many wells in existing networks are in the base-line category, but information about them may be of doubtful value. There are several reasons for this, most of which stem from the fact that many wells used for observation were constructed for some other specific purpose, such as domestic water supply, and were abandoned because of inadequate yield. Many do not respond readily to fluctuation of the water table or of artesian pressure in the main aquifer. Another problem is that data from different wells are not comparable because of differences in topographical position, geological setting and well construction.

It may be useful to think of a base-line network as being composed of subnetworks A and B. Subnetwork A consists only of observation wells that indicate the effects of areal variations in climate on ground-water storage. In other works, wells in this subnetwork must be of nearly identical construction and must be situated in nearly identical geological and topographical conditions. Subnetwork B consists of wells that show how ground-water behaviour is modified by differences in topography and geological setting, when climatic conditions are comparable.

Using Table 7.6.6.2, it is possible to subdivide an area and to select observation stations which take into account the effect of all factors and natural conditions. The subdivisions

TABLE 7.6.6.2. Factors and natural conditions affecting natural ground-water fluctuations

Regions	Ground-water conditions and general characteristics of water-level fluctuations
A. *Climatic*	
1. Permafrost areas	Two summer water-level rises
2. Uniform freezing in the soil zone at the land surface	Marked water-level rise in the spring, followed by water-level recession until autumn. A second smaller water-level rise in autumn, followed by gradual decline until spring
3. Sporadic freezing of the zone of aeration	Water-level rises mainly in the winter
4. Complete absence of soil freezing	Water-level rises during rainy season
B. *Regions with different amounts of moisture*	
1. Region of high moisture	The amount of precipitation is higher than evapotranspiration. Water-levels affected rapidly by small rains and small temperature variations. Small amplitude of water fluctuations
2. Region of moderate moisture	As water table is at greater depth than in zone 1, amplitudes of water level fluctuations are more distinct and greater than in zones 1 and 3
3. Region of small moisture	Evapotranspiration is a dominant factor in water-level fluctuations
C. *Surface drainage and degree of slope*	
1. Well developed drainage (generally mountainous topography)	High runoff and low infiltration to ground water. Water-level fluctuation amplitudes may be high
2. Moderately developed drainage (generally uplands)	Moderate runoff and infiltration to ground water. Water-level fluctuation amplitudes are lower than in zone 1 but higher than in zone 3
3. Poorly developed drainage (generally plains and valley bottoms)	Low runoff and high infiltration to ground water. Water table at shallow depth. High evapotranspiration. Ground-water salinity may be high as compared with zones 1 and 2
D. *Thickness of zone of aeration (d)*	
1. *d* is less than 0.5 m.	Water-level fluctuations of small amplitude. Evapotranspiration from the water table prevails over spring discharge.
2. *d* is between 0.5 and 4 m thick.	Water-level fluctuations of larger amplitude than in zone 1. Spring discharge prevails over evapotranspiration.
3. *d* is greater than 4 m.	Water-level fluctuations are of small amplitude and evapotranspiration might be of limited importance.
E. *Hydrological regions distinguished on the basis of rock types*	
1. Crystalline rocks	The infiltration and rate of movement of the water in the aquifer, etc., depend on the type of rock.
2. Volcanic rocks	
3. Carbonate rocks	
4. Glacial deposits	
5. Sediments compacted and not compacted	
6. Alluvium, etc.	

Source: A. A. Konoplyantsev, V. S. Kovalesky and S. M. Semenov, *(Natural Régimes of Underground Waters and their Characteristics)*, Moscow, Gosgeoltekhizdhat, 231 p. (In Russian.)

should begin by division into climatic provinces. Each climatic province can then be further subdivided into zones on the basis of moisture availability and infiltration probability. In turn, each zone can be subdivided into regions on the basis of drainage development and land slope.

As a result of such subdivision each observation point will be characterized by a definite thickness of the zone of aeration, depending on the particular geological conditions, by a certain type of topography, and must refer to a definite regional moisture content and to a climatic province. With reasonable selection of observation points it is possible to create a key network included in subnetwork A.

Because the wells in subnetwork B are intended to permit evaluation of the effect of both topography and geology on the response of aquifers to climate, wells in this network are placed near some or all of the wells in subnetwork A, but in different topographical positions or deeper aquifers. For example, if the wells in subnetwork A are situated in valleys, some of the wells in subnetwork B would be placed on hilltops and hillsides.

The number of wells needed in a base-line network depends on the number of climatic zones in a region, the diversity of topography, and the complexity of the geohydrological environment. For example, the state of North Carolina, in the United States (area 136,198 km^2), has been divided by the United States Weather Service into eight climatic zones. Subnetwork A therefore should include at least one well in each of these zones. These wells should be sited on relatively level interstream areas remote from major pumping centres, and should be finished with screens placed approximately 2 to 3 m below the lowest seasonal level of the water table.

Subnetwork B should consist of wells penetrating deeper aquifers at some or all of the same sites. In the North Carolina piedmont and mountainous region, underlain by crystalline igneous and metamorphic rocks, subnetwork B need consist only of single wells open to bedrock. In the coastal-plain region, however, underlain by two or more successive aquifers, at least two wells screened in the deeper aquifers would be desirable at some of the locations. Thus, in Noth Carolina the minimum ideal base-line network would consist of eight wells in subnetwork A and possibly as many as twelve wells in subnetwork B, making a total of twenty water-level observation wells in the base-line network of the state.

Because wells in the base-line network provide information on natural fluctuations which is necessary for the analysis of data from the other networks, the wells in subnetwork A should be intended for long-term operation. When the effects of topography and geohydrological environment have been established through correlation of results from subnetworks A and B, further recordings from subnetwork B wells may not be needed.

7.6.6.3 **Water-management networks**

The water-management network consists of wells used to measure how ground-water levels respond to artificial or man-made stresses. This is a localized type of network situated in or near pumping centres, artificial recharge sites, deep waste injection sites, etc. Water-level observation programmes in all developed and in many developing nations undoubtedly include some wells that would meet this criterion.

The number of wells needed differs from region to region, depending on the complexity of the ground-water systems, the extent of local ground-water development, and the number and magnitude of areas of stress. At least one well should be placed near every major pumping centre, in a position where it will measure the composite drawdown rather than reflect the influence of only one nearby well. A major centre is one with a pumping rate high enough to produce drawdowns over a locally extensive area. In the cavernous limestone aquifers of Florida this could mean withdrawals exceeding 100,000 m^3/day. In the glacial-outwash aquifers of New York it might refer to withdrawals exceeding 20,000 m^3/day.

It is probably useful to consider the objectives of the water-management network. The chief objective is to measure the response of ground-water systems to stress, because this is essential to prediction of their long-term yield. A second objective, directly related to the first, is to provide information essential to water management, including warnings of overpumping, and the effects of extending the existing well fields.

Some observation wells at major pumping centres should record water-level variations in the producing aquifer at different distances from the centre of pumping. In order to measure the three-dimensional response of the system to withdrawals, other wells should be screened at different depths, in two or more permeable zones.

In deciding how long the observations should be continued, it is useful to note that most pumping centres belong to one of two groups: (a) those at which the rate of withdrawal remains constant over a long period of years; and (b) those at which the rate of withdrawal is being gradually or intermittently increased. Where withdrawals have been constant, and are likely to remain so, there is no need to continue observations after the drawdowns have stabilized. Where withdrawals are increasing, observations must be continued for an indefinite period.

North Carolina is again an example of the needs for minimum-sized water-management networks. In the coastal-plain region there are possibly as many as ten pumping centres which are important enough to be included in a water-management network. Between fifty and sixty observation wells, some of which are already in operation, would be required to obtain the necessary water-level data for these centres. In the piedmont and mountainous region of North Carolina there are as yet no major pumping centres, although as surface reservoir sites are used up one by one, increasing use will be made of the ground water. For this reason steps are being taken to intensify the collection of water-level data in the piedmont and mountainous region of the state, so as to observe how the ground-water system responds to stress. This programme will include perhaps six or eight wells near two or three of the largest existing pumping centres where withdrawals are most likely to increase in the near future.

7.6.6.4 **Special-purpose networks**

In addition to the water-level networks discussed above, other types of special purpose networks include water-management networks in mining areas, in irrigation projects, near dams and reservoirs, near navigation canals and locks, and in drainage projects.

Networks in mining areas may measure seasonal inflows of surface and ground water into mine galleries or surface pits; establish the hydraulic links between water in mines and adjacent streams or lakes; monitor any potentially dangerous hydrostatic head differences in fault zones or along stratigraphic interfaces that could result in water bursts during mining operations; and study the local geohydrological environment and how it is modified by the extension of surface or underground mine workings (Babuskin *et al.*, 1969).

In irrigation projects, networks are important in evaluating the seasonal and long-term changes in ground-water levels which result from continuing applications of irrigation water; in studying the influence of canals on the ground-water system; in the preparation of periodic water and salt budgets for irrigation projects; in designing conjunctive use irrigation projects involving both surface and ground water; and in evaluating the effects of suitably placed drainage measures in controlling ground-water levels.

Water-level observation wells or piezometers near dams and reservoirs are useful in determining the preconstruction condition of the ground-water system, as well as the response of the system to the stresses of construction. These stresses include the loading effects which filled reservoirs and heavy dams may have on aquifers and confining beds in the ground-water system; the changes in ground-water conditions around and near reservoirs; the influence of reservoir losses from seepage; and the effects of seepage on the stability of a dam's foundation.

The effects which canals and locks have on a ground-water system, either by excavation and loading or from seepage, also require monitoring. The wells are commonly placed along and across canal lines and in the vicinity of locks.

Water-level observations are necessary in waterlogged lands being dewatered by lateral or vertical drainage systems. Water-level measurements are made to ensure that the water table can be kept below a maximum level, and that the drainage system is operating at maximum efficiency and effectiveness (Maslov, 1970; Shestakov, 1965).

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8 Analysis and presentation of the data

8.1 Statistical techniques

The formation of the ground-water régime is a complicated process which depends on a great number of physical characteristics. At times, processes acting simultaneously have opposite effects on certain ground-water properties. For example, precipitation which infiltrates down to the saturated zone tends to raise the ground-water level while, at the same time, evaporation tends to lower the level.

Mathematical statistics can often be used to determine the manner in which various factors affect the ground-water régime. Relations between ground-water characteristics and other variables can often be readily obtained by using simple- or multiple-regression analysis.

8.1.1 *Correlation and regression*

The distinctions between correlation and regression must be recognized in order to apply and interpret either of the methods. These distinctions are very marked although they may seem of little importance because of the similarity of the computational procedures. Dixon and Massey (1957, p. 189) make the following distinction between the two: 'A regression problem considers the frequency distribution of one variable when another is held fixed at each of several levels. A correlation problem considers the joint variation of two measurements, neither of which is restricted by the experiment.'

Correlation is a process by which the degree of association between samples of two variables is defined. The correlation coefficient is a mathematical definition of that association. It is, of course, possible to compute a correlation coefficient from any two sets of data. The mathematical definition of association implies no cause-and-effect relation nor even that the relation between the two variables results from a common cause.

Correlation theory requires that the data be drawn randomly from a bivariate normal distribution. A further requirement of correlation is that both variables X and Y be without error due to measurement. Nothing can be measured without error, so the above requirement is one of degree. The question of the error allowable is subject to arbitrary decisions, particularly since the true error of the data is never known.

The end product of the process of correlation is the correlation coefficient; *it is not an equation.* The equations which describe Y as a function of X, and X as a function of Y, are regression equations, not correlation equations. Another way of stating the distinction between correlation and regression is that correlation measures the degree of association between two variables, whereas regression provides equations for estimating individual values of one variable from given values of the other.

If the data can reasonably be assumed to be drawn from a normal bivariate distribution, then both correlation and regression analyses are appropriate. It is under this assumption that most of the examples in statistics texts are analysed. However, regression is also appropriate under certain other conditions when correlation is not. The only assumptions required for regression are:

1. The deviations of the dependent variable about the regression line (for any fixed X) are normally distributed, and the same variance exists throughout the range of definition.
2. Values of the independent variable are known without error. The dependent variable is considered as an observation on a random variable, and the independent variable as some known constant associated with this random variable.
3. For any value of the independent variable, observed values of the dependent variable are uncorrelated random events.
4. Each of the variables is homogeneous; that is, all individual values of a variable measure the same thing. Data are considered homogeneous if any subgroup to which certain of these data may be logically assigned has the same expected mean and variance as any other subgroup of the population. Neither variable need have a probability distribution in regression (but, of course, Y values corresponding to a fixed X are assumed to be normally distributed).

The regression equation gives the average amount of change in the dependent variable corresponding to a unit change in the independent variable. Thus, it gives more specific information than correlation. The regression coefficient can be tested to determine whether it is significantly different from zero, and this test is identical to the test of significance of the correlation coefficient (provided the data are drawn from a bivariate normal distribution).

The over-all reliability of a regression is measured by the standard error, which is approximately the standard deviation of the distribution (assumed normal) of residuals about the regression line. Figure 8.1.1a shows schematically the distribution of residuals, and the three individual examples have been selected to show the theoretical normal distribution around the regression line at points A, B and C. By definition, the standard error is the same throughout the range of X. This standard error was called the standard error of estimate by Ezekiel (1950, p. 131). It is also referred to as the standard error of regression and as the standard deviation from regression.

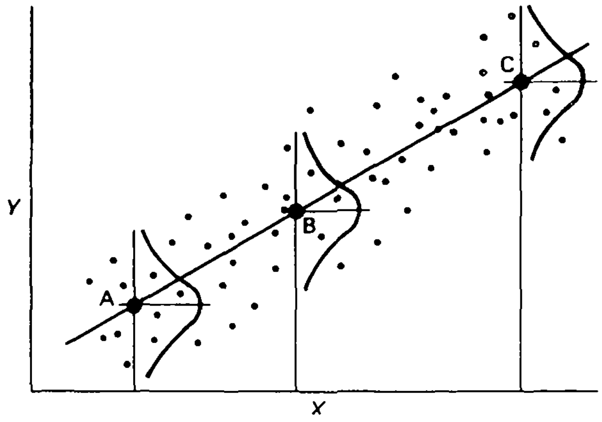

FIG. 8.1.1a. Normal distribution of plotted points about the regression line

The standard error of a prediction from regression is made up of three parts: the error of the mean, the error of the slope of the line and the standard error of estimate. All three may be expressed in terms of the standard error of estimate so that the standard error of a prediction (S_p) is:

$$S_p = S_e \sqrt{1 + \frac{1}{n} + \frac{(X - \bar{X})^2}{\Sigma (X - \bar{X})^2}} \qquad 8(1)$$

where S_e is the standard error of estimate, n is the number of items in the sample, X is the independent variable and $\bar{X}$ is the mean of the X values. Thus the error of a prediction increases with distance from the mean (Snedecor, 1948, p. 120). The standard error of prediction is used to evaluate the error associated with a particular value of the dependent variable as determined from a given value of the independent variable and the regression equation.

Most analyses require use of multiple correlation or regression. A multiple correlation is evaluated by partial correlation coefficients and by an index of total correlation. A partial correlation coefficient is an index of the degree of association between one independent variable and the dependent variable after the effects of the other independent variables have been removed.

In a multiple-regression equation the regression coefficients are called partial regression coefficients. Each shows the effect on Y of a unit change in the particular independent variable, the effects of the other independent variables being held constant.

If the independent variables in a regression analysis are related to each other, the partial regression coefficients will be of a different magnitude than the simple regression coefficients. The independent variables in a regression usually are related to each other as well as to the dependent variable.

The assumptions required for correlation are infrequently met in engineering problems and seldom met in hydrologic problems. Many of these problems to which the correlation method does not apply can be handled by the regression method because of the less restrictive assumptions. Thus the regression method may be used for such relations as that of concrete strength to time of setting, where neither value is randomly selected and neither variable has a probability distribution. Obviously the range of such a relation is limited to the range of the data selected.

Under the above conditions the correlation coefficient does not apply but, of course, can be computed from the relation

$$r = \sqrt{1 - (S_e/S_y)^2} \qquad 8\,(2)$$

where r = correlation coefficient, S_e = standard error of estimate, and S_y = standard deviation of the values of the dependent variable. From the above formula it can be seen that r depends on S_y, which depends on the range of data selected for problems such as the concrete strength relation to time of setting. Therefore, if the variables used in a regression are not randomly sampled, the computed value of r changes with the range of the arbitrarily selected sample and is therefore meaningless. Empirical verification of this statement is given by the data plotted in Figure 8.1.1b. (These data were selected to demonstrate this principle; the relation is not hydrologically significant.) Using all the points, the relation is computed to be:

$$\log Y = 2.27 + 0.59 \log X,$$

the standard error is 0.22 log cycle and the computed correlation coefficient is 0.97.

If only the 14 points for X values ranging from 40 to 2,000 (Fig. 8.1.1.b) are used, the relation is:

$$\log Y = 2.31 + 0.57 \log X.$$

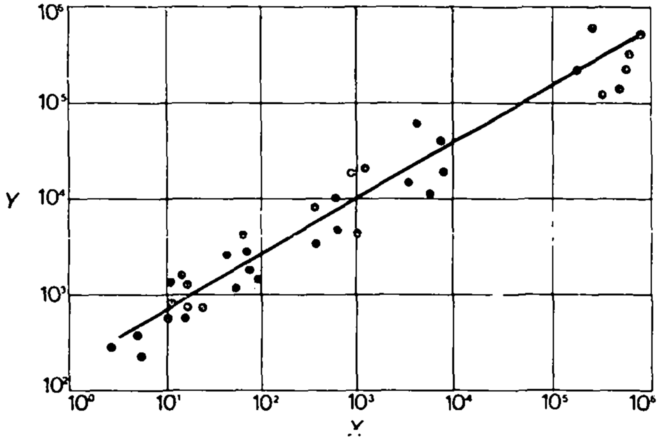

FIG. 8.1.1b. Plot used in demonstrating the effect of sample range on computed correlation coefficient.

This relation has a standard error of 0.23 log cycle (almost the same as the previous standard error), but the computed correlation coefficient is 0.83, much lower than that obtained by using samples from a greater range. Obviously such variability in the correlation coefficient would render it unsuitable as a measure of the degree of relation for this type of application.

Regression is emphasized over correlation, not only because correlation is commonly inapplicable to particular hydrologic data but because regression provides quantitative answers to specific problems. In general, regression is preferred over correlation for hydrologic problems even when the data are suitable for a correlation analysis. Uses of regression analysis are (Konoplyantsev, 1967; Riggs, 1968):

1. To estimate individual values of the dependent variable corresponding to selected values of the independent variables.
2. To determine the amount of change in the dependent variable associated with a unit change in an independent variable.
3. To determine whether certain variables (which do not have probability distributions) are related to a dependent variable.
4. To improve estimates of the parameters defining the probability distribution of the dependent variable.

Correlation is most useful in theoretical studies and in time-series analysis.

8.1.2 *Simple linear regression*

Computation of a regression equation using the simple linear model $Y = a + bX$ is demonstrated using the data given in Table 8.1.2. That table also shows computation of means, and the sums of cross-products, and squares. The individual cross-products and squares need not be recorded; the sum of cross-products, or squares, can be cumulated on a desk calculator. Such calculations are ordinarily checked by repeating the operation.

The coefficients a and b in the regression equation and the standard error of estimate are computed as shown below.

TABLE 8.1.2. Data and computations for example of two-variable regression

Period		Recharge component in metres (Y) (1)	Precipitation in centimetres (X) (2)	XY (3)	X^2 (4)	Y^2 (5)
Oct.-Nov.	1935	1.25	10.36			
Oct.-Nov.	1936	1.40	8.94			
Oct.-Nov.	1937	2.13	13.21			
Oct.-Nov.	1938	1.19	15.80			
Sept.-Oct.	1939	1.65	11.18			
Sept.-Oct.	1940	1.89	13.64			
Oct.-Nov.	1941	1.68	19.53			
Nov.-Dec.	1942	1.77	24.56			
Oct.	1943	1.28	11.48			
Nov.	1944	1.16	7.77			
Aug.	1945	.94	11.30			
Oct.-Dec.	1945	3.69	28.12			
Oct.-Dec.	1946	3.51	30.18			
Aug.-Oct.	1947	3.14	23.14			
Oct.-Nov.	1948	1.22	15.88			
Oct.-Nov.	1949	2.29	19.76			
Sept.-Dec.	1950	4.42	35.36			
Oct.-Dec.	1951	2.90	25.40			
Sum		37.51	325.61	804.1725	7002.0551	96.0677
Mean		2.084	18.089			

Col. (1). Recharge component in the adjusted water-level changes in the Tower Well, Weber County, Utah (United States) (from Thomas, 1963).
Col. (2). Precipitation at Pine View Dam, Utah (United States).

$$b = \frac{\Sigma XY - \dfrac{\Sigma X\,\Sigma Y}{N}}{\Sigma X^2 - \dfrac{(\Sigma X)^2}{N}} \qquad 8(3)$$

where $\bar{X}$ and $\bar{Y}$ are the mean values and N is the number of observations; in this example $N = 18$.

$$b = \frac{804.172 - \dfrac{(325.61)(37.51)}{18}}{7002.06 \quad \dfrac{(325.61)^2}{18}} = 0.113. \quad \textit{Regression coefficient.}$$

$$a = Y - b\bar{X} = 2.084 - (0.113)\,(18.089) = 0.040. \quad \textit{Intercept.}$$

Then

$$Y = a + bX = 0.040 + 0.113\ X,$$

Or

$$Y = \bar{Y} + b\,(X - \bar{X}) = 2.084 + (0.113)\,(X - 18.089),$$

$$Y = 0.040 + 0.113\ X. \quad \textit{Equation of least-squares line.}$$

$$S_x^2 = \frac{\Sigma X^2 - \frac{(\Sigma X)^2}{N}}{N - 1} = \frac{7002.06 - \frac{(325.61)^2}{18}}{17} = 65.409. \quad \textit{Variance of } X.$$

$$S_y^2 = \frac{\Sigma Y^2 - \frac{(\Sigma Y)^2}{N}}{N - 1} = \frac{96.0677 - \frac{(37.51)^2}{18}}{17} = 1.053. \quad \textit{Variance of } Y.$$

$$S_{y.x}^2 = \frac{N-1}{N-2}\left[S_y^2 - b^2 S_x^2\right] = \frac{17}{16}\left[1.053 - (0.113)^2\, 65.409\right] = 0.231.$$

$S_{y.x} = 0.481$. *Standard error of Y.*

$$r = b\,\frac{S_x}{S_y} = 0.113\,\frac{8.09}{1.026} = 0.89. \quad \textit{Correlation coefficient.}$$

The regression coefficient can be tested for significance by means of the student's t test (Bennett and Franklin, 1954, p. 228) for which tabulated values are available (see Appendix G) as follows:

$$S_b^2 = \frac{S_{yx}^2}{\Sigma(x^2)} = \frac{0.2314}{1111.95} = 0.000208$$

where $\Sigma(x^2) = \Sigma(X - \bar{X})^2 = \Sigma X^2 - NX^2$.

Testing the hypothesis that $\beta = 0$, where β is the true value of the slope of the regression line which is estimated by b:

$$t_{N-2,\,\alpha} = \frac{b - \beta}{S_b} = \frac{0.113 - 0}{0.0144} = 7.847.$$

$t_{N-2,\,\alpha}$ refers to the student's t-distribution with $(N - 2)$ degrees of freedom at a significance level of $1 - 2\alpha$.

From a table of t, $t_{16,0.005} = 2.95$. The 99-per-cent confidence limits for β are:

$$b - t_{16,0.005}\, S_b < \beta < b + t_{16,0.005}\, S_b,$$

$$0.113 - 2.92\,(0.0144) < \beta < 0.113 + 2.92\,(0.0144),$$

$$0.0710 < \beta < 0.155.$$

The confidence limits give the range within which β lies with a 99 per cent probability in the foregoing example.

If zero were included in the confidence limits, then it could be said that the computed slope was not significant at the selected significance level.

The locus of the regression equation and the data used are shown in Figure 8.1.2.

8.1.3 Multiple linear regression

A multiple linear regression model has the form:

$$Y = a + b_1X_1 + b_2X_2 + \dots b_nX_n$$

The regression constants for such a model are computed from m normal equations. For a model with two independent variables, that is:

$$Y = a + b_1X_1 + b_2X_2, \tag{8(4)}$$

the normal equations are

$$b_1\, \Sigma\,(x_1^2) + b_2\, \Sigma\,(x_1x_2) = \Sigma\,(y\, x_1),$$

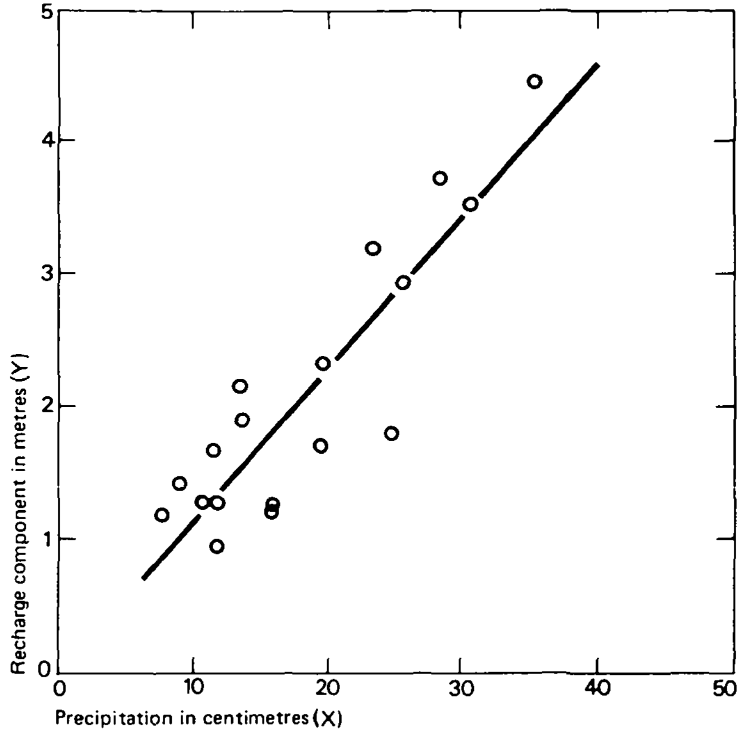

FIG. 8.1.2. Plot of data from Table 8.1.2 and the computed regression line.

$$b_1 \Sigma (x_1 x_2) + b_2 \Sigma (x_2^2) = \Sigma (y\, x_2), \tag{8(5)}$$

and

$$a = \bar{y} - b_1 \bar{X}_1 - b_2 \bar{X}_2. \tag{8(6)}$$

The symbols $\bar{y}$ and $\bar{X}_i$ represent the mean of the dependent variable and the mean of the ith independent variable respectively; X_i represents a particular value of the ith variable, and x_i represents $(X_i - \bar{X}_i)$, the deviation from the mean of that variable. It is simpler to compute the squares and cross-products of the variables in terms of X and then convert the results in terms of x than to begin with deviations from the mean. The conversion equations are

$$\begin{aligned} \Sigma (y x_1) &= \Sigma (Y X_1) - N \bar{Y} \bar{X}_1, \\ \Sigma (x_1^2) &= \Sigma (X_1^2) - N \bar{X}_1^2, \\ \Sigma (y x_2) &= \Sigma (Y X_2) - N \bar{Y} \bar{X}_2, \\ \Sigma (x_1 x_2) &= \Sigma (X_1 X_2) - N \bar{X}_1 \bar{X}_2, \end{aligned} \tag{8(7)}$$

and

$$\Sigma (x_2^2) = \Sigma X_2^2 - N \bar{X}_2^2$$

where the last term in each equation is called the correction item and N is the number of items in the sample.

For three independent variables the model is of the form

$$Y = a + b_1X_1 + b_2X_2 + b_3X_3. \tag{8(8)}$$

The normal equations are

$$\begin{aligned} b_1\,\Sigma\,(x_1^2) + b_2\,\Sigma\,(x_1x_2) + b_3\,\Sigma\,(x_1x_3) &= \Sigma\,(yx_1), \\ b_1\,\Sigma\,(x_1x_2) + b_2\,\Sigma\,(x_2^2) + b_3\,\Sigma\,(x_2x_3) &= \Sigma\,(yx_2), \\ b_1\,\Sigma\,(x_1x_3) + b_2\,\Sigma\,(y_2x_3) + b_3\,\Sigma\,(x_3^2) &= \Sigma\,(yx_3), \end{aligned} \tag{8(9)}$$

and

$$a = \overline{Y} - b_1X_1 - b_2\overline{X}_2 - b_3X_3, \tag{8(10)}$$

where the symbols are the same as before.

The standard error of estimate, S_e, for two independent variables is computed from the equation.

$$S_e^2 = \frac{\Sigma\,(y^2) - b_1\,\Sigma\,(yx_1) - b_2\,\Sigma\,(yx_2)}{N - M} \tag{8(11)}$$

where N is the number of items in the sample and M is the number of lost degrees of freedom (one degree of freedom is lost for each constant in a regression equation).

For three independent variables

$$S_e^2 = \frac{\Sigma\,(y^2) - b_1\,\Sigma\,(yx_1) - b_2\,\Sigma\,(yx_2) - b_3\,\Sigma\,(yx_3)}{N - M} \tag{8(12)}$$

The method of computation is best described by use of an example. The model, the data, and the preliminary computations are shown in Table 8.1.3. Only the cumulative sums of cross-products and squares are taken from the calculator and recorded in Table 8.1.3; individual values are not needed. The correction items shown in Table 8.1.3. are obtained from the last term in the appropriate conversion equation. For x_1x_2 the appropriate equation is

$$\Sigma x_1x_2 = \Sigma X_1X_2 - N\overline{X}_1\overline{X}_2$$

and the correction item of Table 8.1.3 is $N\overline{X}_1\overline{X}_2$.

Subtracting the correction item from ΣX_1X_2 gives Σx_1x_2 which is the corrected sum in Table 8.1.3. This and the other corrected sums are then substituted in the normal equations. The computation of regression coefficients is shown below with the explanation following.

TABLE 8.1.3 Multiple regression example: Baldwin Park, California, observation well

Year	Maximum April-June ground-water level in metres above msl (Y)	Minimum prior October-December ground-water level in metres above msl (X_1)	January-March precipitation in centimetres (X_2)	Prior October-December precipitation in centimetres (X_3)
1941	97.17	87.29	57.78	24.13
1942	94.88	95.37	7.62	17.83
1943	99.03	89.98	52.88	5.99
1944	98.94	93.67	30.48	19.79
1945	95.86	94.98	20.80	18.26
1946	93.57	92.60	14.25	17.63
1947	94.00	91.32	7.32	25.91
1948	88.24	88.88	12.22	8.79
1949	84.95	84.83	18.47	9.83
1950	81.84	81.11	16.99	13.94
1951	78.76	78.61	11.18	6.55
1952	83.85	74.80	43.03	22.78
1953	82.91	79.49	6.83	19.86
1954	79.58	77.18	33.20	5.49
1955	78.09	76.96	17.50	8.99
1956	76.72	75.19	23.27	6.63
1957	75.86	74.43	24.00	2.24
1958	86.47	72.85	30.86	26.80
1959	81.90	81.93	15.98	1.52
1960	74.68	76.29	16.05	5.74
Sum	1 727.30	1 667.76	460.71	268.70
Mean	86.3650 $(\bar{Y})$	83.3388 $(\bar{X}_1)$	23.0355 $(\bar{X}_2)$	13.4350 $(\bar{X}_3)$

Model is $Y = a + b_1X_1 + b_2X_2 + b_3X_3$. N = 20.

	X_1X_2	X_1X_3	YX_1	X_2^2
Sums	38 245.5533	22 771.3964	145 076.8616	14 604.8395
Correction terms	38 417.6855	22 406.3556	144 036.0924	10 612.6852
Corrected sums	− 172.1322	365.0408	1 040.7692	3 992.1543
	X_2X_3	YX_2	X_3^2	YX_3
Sums	6 476.7515	40 488.2504	4 881.1624	23 935.3793
Correction terms	6 189.6388	39 789.2192	3 609.9845	23 206.2755
Corrected sums	287.1127	699.0312	1 271.1779	729.1038
	Y^2	X_1^2		
Sums	150 462.8820	140 201.3672		
Correction terms	149 178.2645	139 071.1708		
Corrected sums	1 284.6175	1 130.1964		

Normal equations (see Ezekiel, 1950, p. 198):

$$
\begin{array}{lrcl}
\text{I} & b_1\,\Sigma\,(x_1^2) + b_2\,\Sigma\,(x_1x_2) + b_3\,\Sigma\,(x_1x_3) & = & \Sigma\,(yx_1) \\
\text{II} & b_1\,\Sigma\,(x_1x_2) + b_2\,\Sigma\,(x_2^2) + b_3\,\Sigma\,(x_2x_3) & = & \Sigma\,(yx_2) \\
\text{III} & b_1\,\Sigma\,(x_1x_3) + b_2\,\Sigma\,(x_2x_3) + b_3\,\Sigma\,(x_3^2) & = & \Sigma\,(yx_3) \\
\text{I} & 1{,}130.1964b_1 - 172.1322b_2 + 365.0408b_3 & = & 1040.7692 \\
\text{I}' & b_1 - 0.152303b_2 + 0.322989b_3 & = & 0.920875 \\
\text{II} & -172.1322b_1 + 3992.1543b_2 + 287.1127b_3 & = & 699.0312 \\
\text{I} & 172.1322b_1 - 26.2163b_2 + 55.5968b_3 & = & 158.5122 \\
\Sigma_2 & 3965.9380b_2 + 342.7095b_3 & = & 857.5434 \\
\text{II}' & -b_2 - 0.0864132b_3 & = & 0.216227 \\
\text{III} & 365.0408b_1 + 287.1127b_2 + 1271.1779b_3 & = & 729.1038 \\
\text{I} & -365.0408b_1 + 55.5968b_2 - 117.9042b_3 & = & 336.1569 \\
\Sigma_2^1 & -342.7094b_2 - 29.6146b_3 & = & -74.1031 \\
\Sigma_3 & 1123.6591b_3 & = & 318.8438 \\
 & b_3 & = & 0.2838 \\
\text{II}' & b_2 - 0.0864132(0.2838) & = & 0.216227 \\
 & b_2 & = & 0.1917 \\
\text{I}' & b_1 - 0.152303(.1917) + 0.322989(0.2838) & = & 0.920875 \\
 & b_1 & = & 0.8584 \\
\text{III} & 365.0408(0.8584) + 287.1127(0.1917) + 1271.1779(0.2838) & = & 729.1038 \\
 & 729.1508 & \simeq & 729.1038.\ \textit{Check.}
\end{array}
$$

The regression constant is obtained from

$$a = \bar{Y} - b_1\bar{X}_1 - b_2\bar{X}_2 - b_3\bar{X}_3$$

$$a = 86.365 - (0.8584)(83.3388) - (0.1917)(23.0355) - (0.2838)(13.435)$$

$$a = 6.598$$

Substituting the computed constants in the regression model gives

$$Y = 6.60 + 0.858\,X_1 + 0.192\,X_2 + 0.284\,X_3.$$

The standard error of estimate, S_e, is computed as follows:

$$S_e^2 = \frac{\Sigma\,(y^2) - b_1\,\Sigma\,(yx_1) - b_2\,\Sigma\,(yx_2) - b_3\,\Sigma\,(yx_3)}{N - M}$$

$$= \frac{1284.6175 - 0.8584(1040.7692) - 0.1917(699.0312) - 0.2838(729.103)}{20 - 4}$$

$$= \frac{50.2973}{16} = 3.1436.$$

$$S_e = 1.772$$

= standard error in metres.

The above computation utilizes the Doolittle method, a simplified method of solving simultaneous equations having a certain symmetry. The normal equations are on the first three lines. Next is the first normal equation with converted sums from Table 8.1.3 substituted in it. Line 5 is obtained by dividing the equation next above by its coefficient of b_1. Line 6 is the second normal equation, with converted sums from Table 8.1.3 substituted in it. Line 7 is obtained by multiplying the equation of line 4 by the coefficient of b_2 in line 5 with the sign changed. Line 8 is obtained by adding line 7 and line 6. Line 9 is line 8 divided by the coefficient of b_2 with the sign changed. Line 10 is the third normal equation. Line 11 is line 4 multiplied by the coefficient of b_3 on line 5 with the sign changed. Line 12 is line 8 multiplied by the coefficient of b_3 on line 9. Line 13 is the sum of lines 10, 11, and 12. Lines 14-18 complete the computations of the regression coefficients. Lines 19 and 20 are used to check the results. Only the third normal equation provides a complete check.

Methods of solving regression equations involving more than three independent variables are more complex and are discussed by Fisher (1950), Bogardi (1953, 1954), Bennett and Franklin (1954), Remson and Randolph (1958), Lebedev (1957, 1964), Ezekiel and Fox (1959), Thomas (1963), Rogovskaya and Morozov (1964), Konoplyantsev (1967) and Riggs (1968).

The multiple correlation coefficient, R, can be computed by

$$R = \sqrt{1 - \frac{S_e^2}{S_y^2}} \qquad 8(13)$$

where S_y^2 is the variance of the dependent variable and is equal to $\Sigma y^2/(N-1)$ and S_e is the standard error of the multiple regression equation.

The coefficient of determination, R^2, is a measure of the degree to which the variance of the independent variable is accounted for by the dependent variable if the assumptions concerning correlation are met. If $R = 0.80$ then $R^2 = 0.64$ and it can be said that about 64 per cent of the variance of the dependent variable is accounted for by the independent variables. Seldom are the assumptions for correlation analysis met in hydrologic problems, therefore, the standard error of estimate should be used to evaluate the usefulness of the computed equation.

A relative measure of the contribution of each independent variable towards accounting for the variance of the dependent variable can be obtained by the use of β coefficients, which are computed as follows:

$$\beta_1 = b_1 \frac{S_{x1}}{S_y}, \quad \beta_2 = b_2 \frac{S_{x2}}{S_y}, \quad \ldots, \quad \beta_m = b_m \frac{S_{xm}}{S_y} \qquad 8(14)$$

where:

$S_{x_m} = \Sigma x_m^2/(N-1)$ and is the variance of X_m;

S_y = variance of Y; and

b_m = regression coefficient.

The values of β vary between -1 and $+1$. The larger the absolute value of β, the more important is the corresponding independent variable to the relation.

In order to determine the contribution of the least significant independent variable to the relation, the variable having the smallest β coefficient can be eliminated from the equation and a new equation and its standard error computed. A comparison of the standard error with and without the dropped variable can be made. If there is little difference, the variable can be dropped without significantly reducing the reliability of the equation. The process can be carried on until only one independent variable remains.

Computations using more than one or two independent variables and/or using a large sample size are very time-consuming. Digital electronic computers are very well suited for making regression computations, and the trend is for greater use of the digital computer for determining regression equations and for making significant computations.

8.1.4 *Graphical simple regression*

The assumptions required of graphical regression are the same as those required for analytical regression. The results of a graphical regression can be expressed mathematically if no restrictions are added to the graphical analysis, and the standard error can be estimated.

Graphical regression is less restrictive than analytical regression in that the model need not be completely specified in advance. In fact if an analytical model cannot be selected on a physical basis, it is conventional to prepare a preliminary graphical regression which will indicate an appropriate model. For example, consider the four data plots of Fig. 8.1.4a. The first (upper left of Figure 8.1.4a) indicates use of the model

$$Y = a + bX.$$

The second (upper right) requires

$$Y = a + bX + b_1X^2$$

where the direction of curvature determines the sign of b_1. The third plot (lower left) indicates the need for a transformation unless the divergence can be explained by an additional variable. The fourth plot (lower right) shows no relation between Y and X, and, if only a two-variable relation is being considered, no further analysis would be made. A relation between Y and X in the fourth plot may be obscured by the effect of another variable Z which has not been included.

The preparation of simple linear relations between two variables is well known. The regression line is not necessarily the same line as one would draw through the plotted points. There are two regression lines, one for $Y = f(X)$, and another for $X = f(Y)$ (Fig. 8.1.4b). The structural line, which balances the plotted points in both directions, has a slope approximately midway between the two regression lines. The differences in

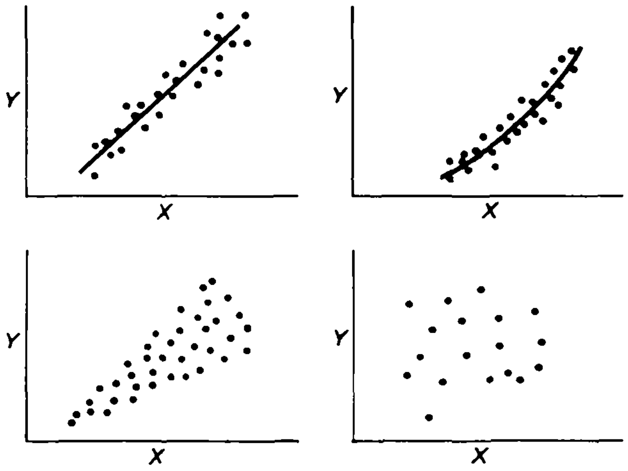

FIG. 8.1.4a. Four possible outcomes of plotting Y against X.

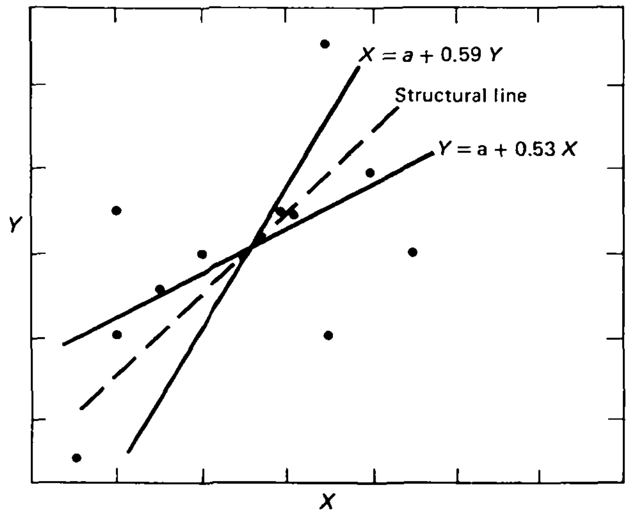

FIG. 8.1.4b. Plot showing the two regression lines and the structural line.

slope among the three lines depend on the degree of correlation of the variables. For perfect correlation all three lines have the same slope. Regardless of the correlation, both regression lines pass through the mean; the structural line may or may not pass through the mean.

To approximate the regression $Y = f(X)$, first, group the points by small increments of X; second, estimate the mean of each group in the Y direction; and third, draw a line which averages these means. It should be kept in mind that the ditribution of points about the regression line in the Y direction is assumed to be the same throughout the range. Obviously that assumption cannot be true for a small number of points, but it is the condition which we try to approximate. The regression line of $Y = f(X)$ will have a flatter slope than that of a line drawn to balance the points in both Y and X directions.

The standard error of estimate of a graphical regression can be readily estimated. The requirements are that (a) the standard error of estimate is essentially the standard deviation of plotted points about the regression line, (b) that about two-thirds of the points should lie between one standard deviation on each side of the mean of a normal distribution, and (c) that a regression line theoretically passes through the mean value of Y corresponding to any value of X. Two lines parallel to the regression line and one standard deviation above and below (in the Y direction), should encompass two-thirds of the plotted points.

The correlation coefficient may also be estimated from a graphical regression by the relation

$$r = \sqrt{1 - \frac{S_e^2}{S_y^2}} \qquad 8(15)$$

where S_e is the graphically determined standard error and S_y is the standard deviation of the Y variables about their mean determined in the same manner as the standard error. Obviously, the correlation coefficient should be estimated only for relations between variables which can reasonably be assumed to be drawn from a bivariate normal distribution.

8.2

Presentation of data on maps

It is expedient, while studying the ground-water régime in large areas, to compile a number of summary maps that permit interpretation of the observed results for the entire territory. The scale of maps depends on the complexity and character of natural conditions in the basins and the accuracy required for the problem solution.

Two groups of summary maps are discussed: (a) analytical maps; (b) synthesizing maps.

The first group of maps are intended to facilitate analysis of the ground-water régime and balance and to present factual material: depths to water, ground-water contours, changes in separate régime elements (water level, chemical composition of water, etc.) in respect of both time and space.

The second group of maps present the interpretative results, or syntheses, of a study. Typical maps would include those that partition a basin according to the types of the ground-water régime and balance formation, the main elements of this balance (percolation of precipitation, evaporation from the water table, local recharge of ground water), values of ground-water recharge to streams, etc.

The contents of such maps and the main principles for their compilation are as follows. Maps that present factual material are compiled on the topographic base chosen for the area. Their legends, or explanations, should permit differentiation to be made between observation wells, water-supply wells, springs, the age and genetic type of water-bearing formations, gauging and measuring stations on streams and lakes, previously drilled test wells, various local artificial and natural factors of the unconfined ground-water régime (wind breaks, irrigated lands, drainage systems), etc. Near the symbols on the map such information as the following can be included: the beginning of the period of observations, depth to water, total dissolved solids content of the water and other background information.

Maps showing depths to the water table are usually compiled for the periods of the maximum, minimum and mean annual position of the water table. The gradations of depths to water should, if possible, be fractional (for instance, from 0 to 1 m, 1-2 m, 2-4 m, 4-6 m, 6-9 m, and so on).

Maps that show amplitudes of ground-water-level fluctuation and the maximum position of the water table are of considerable practical interest. In irrigated areas the above maps are the main documents for appraisal of the extent to which land can be reclaimed. Under natural and disturbed conditions they are used for extrapolating the amounts of ground-water infiltration and evaporation over the entire area. The geological and lithological characteristics of the rocks in the zone of aeration should be shown on the same maps as well as surface factors (vegetation, meso- and microrelief, etc.). Such maps are compiled using all the available data on observation wells, water-supply wells, springs, etc.

Maps showing ground-water contours are compiled when there is a sufficient number of observation wells to determine the main forms of the water table. The number of observation points needed in a particular area is dependent upon the complexity of the area. While compiling a water-table contour map, data on surface-water levels in reservoirs and streams that have free connexion with the unconfined ground-water body should be used.

From maps of ground-water contours compiled on a large, or even small scale, estimates of ground-water flow are made; the degree of heterogeneity of water-bearing formations is appraised and the hydraulic connexion of ground waters with rivers, canals, etc. is determined.

A map of ground-water contours is compiled commonly by linear interpolation between water levels at observation points and by taking the levels of ground water at natural outcrops into consideration. The contour interval chosen depends on the map scale and complexity of the water table. Data on the nature of the hydraulic connexion of ground water with surface waters (usually determined in the field) are also used. The effect of surface relief is carefully analysed and, if possible, the physical characteristics of the water-bearing rocks taken into account, because gradients and even the direction of ground-water flow may be dependent on such factors.

Maps of ground-water contours constructed for various dates are compared by overlaying maps and the level change ΔH for a time period Δt at the points of intersection of ground-water contours is calculated. Maps showing changes in ground-water level for certain periods of time thus may be compiled. The data provided by such maps permit the calculation of the change in ground-water storage ($\mu \Delta H$) and the mapping of the calculated change.

Maps showing the total dissolved solids content of ground waters (in mg/l or other ratio units) or the content of a certain chemical component (e.g. chloride or sulphate ions) are similarly compiled as isolines or by range of concentrations.

The second group of maps, the synthesizing maps, include those which show infiltration, evaporation, local recharge of ground water and discharge into rivers.

A map of infiltration from precipitation is compiled using calculated infiltration at sites where there are at least two observation wells. Infiltration values ($I\Delta\tau$) for each location, expressed as a water layer in mm for some time interval, usually a year, are plotted on the map. If sufficient data are available, linear interpolation, or contouring, of these values is possible. If the amount of such data is small, values can be extrapolated if relationships between infiltration values and depths to water can be established. Infiltration can be expressed either as a water layer in mm per year or in terms of the modulus of infiltration (l/sec/km^2).

Maps of ground-water discharge by evapotranspiration or by film-capillary flow of moisture into the zone of aeration can also be compiled.

From a comparison of maps that show values of infiltration $I\Delta\tau$ and evaporation of ground water $u\Delta V$, the value of unconfined ground-water replenishment or recharge $w\Delta t$ from above may be determined by use of the following relationship: $w\Delta t = I\Delta\tau - u\Delta V$. Maps showing the recharge, $w\Delta t$, are then prepared. By matching maps that show changes in unconfined ground-water storage, $\mu\Delta H$, with maps that show ground-water replenishment, or recharge, from above, $w\Delta t$, maps can be prepared to show areal distribution of the difference between inflow Q_1 and outflow Q_2 of ground waters from the relationship: $[(Q_1 - Q_2)/F]\ \Delta t = \mu\Delta H - w\Delta t$ as a water layer in mm. This difference is usually negative in ground-water recharge areas and positive in ground-water discharge areas.

A weighted average for the value of $[(Q_1 - Q_2)/F]\ \Delta t$ corresponds to an average layer of ground-water flow for the basin for the analysed period (e.g. a year). This can be expressed as a water layer in mm or converted to the infiltration modulus (l/sec/km^2). Consideration of such maps along with a study of the relief, character of vegetation and surface-moistening conditions makes it possible to delineate areas of significant ground-water recharge and discharge and their relation to various natural factors.

Ground-water flow may be appraised on a regional basis and mapped on a small scale by means of stream hydrograph separation. Such mapping assumes that the average value of ground-water discharge to the streams corresponds to an average ground-water outflow from the basin.

Examples of ground-water maps are:

1. Maps of ground-water flow as isolines of mean long-term infiltration moduli (l/sec/km^2) and of coefficients of ground-water flow (as percentage of precipitation).
2. Maps of isolines of minimum ground-water flow moduli (l/sec/km^2).

3. Maps of the average long-term layer of ground-water flow as a water layer (in mm for a period of one year).
4. Maps of ground-water flow expressed as a percentage of the total river run-off.

The first type of map mentioned above shows the distribution of mean long-term, or the absolute, values of discharging ground waters over a basin and the proportion of the total precipitation which replenishes ground water. A hydrogeological map that partitions the basin into recharge and discharge areas is a good base for compiling such maps. The basin can be subdivided further according to aquifer yields. Quality of water information may also be included on such maps.

The map on which the minimum moduli of ground-water flow are shown as isolines is constructed by the use of computed data on minimum values of ground-water discharge. Thus the data on such a map represent the most probable, lowest yearly values of ground-water discharge.

Maps showing ground-water flow as an average long-term layer of water may be used to determine the amount of ground-water recharge. Contours are constructed after the boundaries of each interval of values related to a water layer are determined by consideration of hydrogeological conditions, lithological composition of rocks and geomorphological structure of the basin.

In the compilation of all the above-mentioned maps all the natural factors that influence conditions of ground-water occurrence in the area under investigation must be taken into account. For example, when compiling maps of ground-water flow one should make use of maps or data on such elements of the water balance as ratios of stream discharges, precipitation, evaporation and infiltration. The usefulness of all interpretive hydrogeological maps is additive; that is, their conjunctive usage allows greater knowledge and understanding of an area than when the maps are used separately.

8.3 Forecasting changes in the ground-water régime

The ground-water régime is commonly influenced by both natural and artificial factors. Predictions of the influence of artificial factors on the system can be made only if the natural environmental factors have first been defined. Thus, the forecasting process is complex, involving firstly the definition of how the ground-water régime changes under the influence of natural factors, then the influence of artificial factors. The predicted conditions are thus an expression of the influence of all factors on the dynamics of the ground-water system.

Forecasts of changes in the ground-water régime can be made with the help of: (a) mathematical methods based on variation statistics; (b) hydrodynamic methods based on analytical and numerical solutions to the differential equations of unsteady flow; (c) the water-balance method.

8.3.1 *The statistical method*

This method can be applied in making predictions of changes in a disturbed régime if the number and arrangement of artificial factors remain unchanged over a long-enough period of time (see Sect. 8.1.)

The main criterion for selecting observation points to be used in forecasting is their representative character, with respect to both the geological environment and the geographic locality.

The reliability of this method of prediction depends on the proper choice of independent variables entering the regression equation. An equation or suite of equations for making a prediction may be found after choosing the independent variables. Usually, several versions of the equations need to be worked out with various combinations of factors. The selection of factors is accomplished by taking into account the main genetic features of the ground-water régime typical of each part of the territory being investigated. Meteorological information must be at the disposal of the investigator as this forms the main tool used in making a ground-water forecast. However, it is expedient to use other factors, such as the moisture content in the zone of aeration at various periods and ground-water levels for prior periods, to aid in evaluating the system. The multiple correlation method is the preferred statistical method because this takes into account the influence of a number of interacting factors on the ground-water system.

The controlling environmental factors and the times when these factors exert the most influence upon the ground-water system are rather diverse; therefore, the investigator's approach to the identification of such factors will vary with each problem area. Successful application of analytical methods depends on the experience of the investigator and the ingenuity with which he examines the genetic relationships between the ground-water régime and the natural factors. By applying the multiple correlation method, forecasts of maximum, minimum and average annual ground-water levels can be made. The method may be applied broadly under disturbed conditions of the ground-water régime for studying irrigated areas and for identifying the principal factors affecting the nature of the régime. Electronic computation not only facilitates wider application of the multiple-correlation method in the foregoing types of analytical work but also assists in filling gaps in the record of observations and reconstructing original ground-water levels that preceded disturbance by man's activities. As with most forecasting methods, the shorter the prediction period the more reliable the results become.

A forecast can be made using only observed values of the pertinent factors (for example, precipitation or temperature) before the desired period of prediction, or by using a combination of the observed and forecast data for related factors. For instance, in predicting the spring maximum average monthly ground-water level the forecast can be based solely on water-level data for the immediate past period of record. However, another variant of the forecast is possible by making use of prediction data on precipitation, air temperature, etc., for winter and spring months.

The deviation of the predicted water level from the observed level will be commonly within admissible limits, i.e. less than the mean error. If only a small number of years of record are available the multiple correlation coefficient is overestimated. The best results will be obtained using observation points which have the longest records. A corrected correlation coefficient can be calculated if it is necessary to use short-term records.

The results of a water-level forecast obtained at one representative point can be extrapolated outwards. To effect such extrapolation, the characteristics of the ground-water régime must be recognized. Then the extent to which the known ground-water data can be applied in the unknown territory can be assessed. Correlation between observations at individual points should also be investigated to compile estimated data for any number of points and thus make areal forecasts.

8.3.2 *The hydrodynamic method*

The prediction of ground-water levels, using hydrodynamic equations, invokes an analytical procedure the reverse of that used in determining recharge from observations of the ground-water system.

In order to predict ground-water levels in the case of one-dimensional flow in a semi-infinite water-table aquifer bounded by canals, rivers or reservoirs, it is possible commonly to use an analytical solution for the differential equation describing unsteady ground-water flow. A preliminary estimate of recharge, $w\Delta t$, to the ground-water system moving downward through the unsaturated zone is necessary. This is obtained by analysing the ground-water régime at several locations and by computing the relationship between $w\Delta t$ and such factors as thickness of the zone of aeration and the application of irrigation water. Using these correlations, the expected ground-water recharge for planned conditions of irrigation and drainage can be defined. In addition, the following conditions must be known in order to make water-level predictions: the initial distribution of water levels along a well line; graphs of water-level fluctuations at the boundaries of the ground-water system; the hydraulic diffusivity, a, of the aquifer and the specific yield of the water-bearing materials.

Finite-difference techniques can be applied conveniently in the case of two-dimensional flow. In one such technique the flow area is divided into squares (Fig. 8.3.2) with side length $\Delta x = 2\sqrt{a\Delta t}$ where Δt is a time increment chosen arbitrarily as $t/8$, with t representing the time period over which the forecast is desired.

The head at the node point 0 is calculated using the following expression

$$H_{0,t+\Delta t} = \sum_{i=1}^{4} \frac{H_{i,t}}{4} + \frac{w}{\mu}\Delta t \qquad 8(16)$$

where $H_{i,t}$ is the hydraulic head at node i and time t.

Equation 8(16) is solved for each node of the finite difference grid. Along the boundaries the flow is known in advance and at the remaining nodes the initial head values are determined from the contour map of ground-water levels. The head values determined at time $t + \Delta t$ are used as initial conditions for computing the head at $t + 2\Delta t$ and this procedure is repeated throughout the period of analysis. In practice the very large number of computations involved in this type of analysis requires the use of a digital computer.

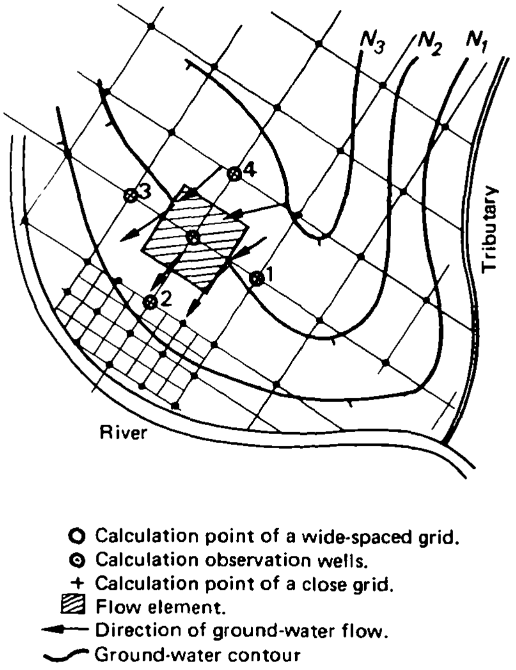

FIG. 8.3.2. Diagram showing location in observation-well nodes in analysing two-dimensional flow.

8.3.3 *The water-balance method*

This method is widely used for experimental determination of all or the majority of water-balance elements. Equation 5.7(2) is used for making computations of the water-balance for areas under irrigation. Ground-water level change (H) over a period of time (t) is the unknown value. This change in level is a weighted average over the irrigated area. The application of this method to irrigated land tracts is discussed.

If lysimeter observations of infiltration of surface waters and precipitation and observations of ground-water evaporation are used, then instead of X, Y_n, K_1, and Z in Equation 5.7(2) the following values are taken into account: $X\psi$ (amount of precipitation which percolates down to the ground-water table, expressed in mm); f_n (percolation of irrigation water supplied to the fields, in cubic metres); K_1 (condensation recharge of free ground waters in mm); U (evapotranspiration from the water table in mm). The ground-water balance is compiled excluding values W_1 and W_2.

The procedure for prediction of ground-water level, using the balance equation, is discussed in the following paragraphs.

The value of precipitation infiltration ($X\psi$) is determined by its dependence on the normal, maximum or minimum amount of precipitation. Such dependences are determined by means of preliminary observations of water-balance elements on similar (zonal and régime) observation stations. The dependence of the infiltration values on the amount of precipitation can be determined with the help of lysimeters.

If Equation 5.7(2) is used for making a forecast, then the normal, long-term maximum or minimum precipitation value (X) is determined from near-by meteorological stations. Values for precipitation are used which correspond to the low, median and high rainfall measurements and thus any prediction will be valid for the corresponding precipitation years, if antecedent conditions are properly taken into account. It is possible to take any probability of occurrence of precipitation and use of this value will then predetermine the corresponding probability of the water-level being predicted.

The probability of a given evapotranspiration rate is also determined by the probability of a given value of precipitation. Similarly the total evaporation (U) from the water surface is dependent on the probability of a given precipitation rate. Such dependences of ground-water evaporation values on the influencing factors are commonly obtained at hydrogeological or hydrometeorological stations.

Condensation of water vapour (K_1) is usually associated with evaporation, values of which are obtained after subtracting the condensation.

The amount of infiltrating water (f_k) from irrigation canals depends on the aquifer parameters and hydrogeological conditions in the irrigated areas. Values of infiltration must be calculated according to the formulae of ground-water dynamics or empirical relationships or determined experimentally at hydrogeological or agricultural (experimental-reclamation) stations.

The amount of water applied for irrigation, (Y_1), for a given irrigation system, is dependent upon the balance element (f_k). This dependency reveals a close relationship between water-level forecasting and the process of designing elements of an irrigation system. The amount of water discharging from an irrigated area (Y_2) is determined by calculation.

The inflowing (Q_1) and outflowing (Q_2) quantities of free ground water are determined by formulae of ground-water dynamics taking account of the hydrological parameters and boundary conditions along the circumference of the irrigated territory.

A change in water reserves on the surface (W_1) can be determined from a preliminary calculated water balance on the surface, especially if the project involves accumulation of surface water in reservoirs.

Changes in moisture content in the zone of aeration can be neglected when the moisture content (at the beginning and end of a defined period) of the zone of aeration corresponds to the lowest moisture capacity. Monthly changes in moisture content of the zone of aeration (W_2) are appraised on the basis of the auxiliary water balances of the zone of aeration and root zone taking into account moisture loss by transpiration. By substitution of the initial data on water-balance elements in Equation 5.7(2) an average change in level (ΔH) over a certain period of time can be determined directly, Commonly, monthly forecasts are made.

A preliminary balance can be assumed to have been compiled satisfactorily if the estimated value of level change (ΔH) during the estimated period (Δt) differs from the observed level by only 10-20 per cent. Before predicting levels by means of the balance method, the total water balance and the unconfined ground-water balance must be analysed for the existing conditions.

8.3.4 The analogy method

In old irrigated areas where the principal ground-water régime factors can be readily identified, advantage may be taken of studies already accomplished concerning definition of the interrelations between the principal elements of the ground-water balance. Thus close correlations may already be available between the value of local replenishment of ground-water flow $(\Delta Q/F)\Delta t = [(Q_1 - Q_2)/F]\Delta t$ and the value of ground-water recharge, ($w\Delta t$), determined for the same time periods. Of course, definite relationships between these values are unique to various geomorphological and hydrogeological regions.

Regional values of $w\Delta t$ and $[(Q_1 - Q_2)/F]\Delta t$ can be calculated from experimental relationships such as $\Delta Q/F = f(w)$ or from the interdependence of the coefficient of recharge $v = \Sigma w \Delta t / \Sigma (Y_n + X) \Delta t$ and the coefficient of land usage (the percentage area irrigated). Knowledge of the preceding relationships in turn helps in determining an average (for the region) change in water level according to the formula:

$$\Delta H_{av} = \left(\frac{Q_1 - Q_2}{F} + w \right) \frac{\Delta t}{\mu} \qquad 8(17)$$

In order to apply this method it is necessary to: (a) calculate recharge of ground water $w\Delta t$ for a reconstructed irrigation area from data obtained along well lines; (b) carry out ground-water balance calculations for separate sites; (c) compare the value of inflow (after subtraction of outflow) $\Delta Q/F$ with the recharge rate w; (d) compile maps of average annual unit ground-water discharges for defined widths of aquifer and compare isolines showing these discharges with a map of ground-water contours to subdivide the area according to the conditions of ground-water flow and balance formation.

A somewhat modified analogous method is applied for prediction of ground-water level changes on newly irrigated lands. Water balances for the zones of aeration and of ground-water flow are compiled from data on infiltration through the soil surface, $w_a \Delta t$, at a site comparable to and from data on the planned water application at the given site.

The following are also taken into account: percentage of infiltration of irrigation water and precipitation on the analogous site; moisture deficit in the zone of aeration up to maximum field moisture capacity at the beginning and the end of the year; difference between inflow and outflow of unconfined ground-water under the site (Lebedev, 1957).

The best methods for making predictions are simulation methods used also for ground-water régime analysis (Bochever *et al.*, 1965). Further consideration of this method will not be given here except to make the following comment: to apply this method a careful study of the natural conditions and the water-economy situation should be undertaken so that the calculated parameters and boundary conditions will most fully reflect the conditions of ground-water movement and the designed scheme of irrigation or drainage.

8.4 Compilation of results of ground-water investigations

8.4.1 *Forms and arrangement of annual information on ground-water régime and balance*

The second session of the IHD Co-ordinating Council (Paris, 19-25 May 1966) recommended that the Unesco Member States publish annual, monthly and typical data on standard observations carried out in representative and experimental basins. The data on a number of hydrometeorological elements including the ground-water régime, accumulated during each year, should be published in the following year.

Annual data on the ground-water régime, obtained at IHD national stations, should be presented as a yearly summary of factual material accompanied by a brief descriptive outline of the régime. Such an outline of the ground-water régime is compiled on the basis of regular and systematic analysis of the observations. This outline is later compared with the information obtained for previous years. The results of ground-water balance calculations and conclusions on other problems being studied at the station which may pertain to or influence the ground-water régime should be used in the compilation of such an outline.

Countries covering large areas with an extensive network of ground-water observation points should group the annual data into defined zones. Each group of data should relate to separate IHD stations and basins and should include the following sections: (a) introduction; (b) descriptive outline of ground-water régime; (c) data on ground-water régime (tables). The data can be arranged according to their geographical location, moving from west to east or from north to south.

The foreword should contain information on the number of stations and representative basins for which observation materials are available, a diagram of their distribution on a map of the country (10 × 15 cm in size), purposes of each station and basin, data on the scope, methods and terms of observation and instrumentation, explanation of abbreviations, legends, etc.

When describing each station or representative basin the introduction should include:

1. Data on geographical location of the station or basin giving co-ordinates of the centre or boundaries of the territory; purpose of and specific tasks to be undertaken at the station or basin (for instance, investigation of the influence of reforestation on the ground-water régime; investigation of ground-water régime and recharge in permafrost areas, etc.). Information on the total number of observation points and their type should be presented in tabular form.
2. An areal map showing location of lines of observation points, the nearest meteorological and gauging stations and the principal data on geology and geomorphology.
3. Hydrogeological sections of the area showing the geological age and lithological composition of rocks and maximum and minimum observed levels of ground waters.

The outline of the ground-water régime should be brief and based on the analysis of the data for the reported year. If there are yearly or seasonal calculations of the ground-water balance, the results of such calculations are to be used when compiling the outline. The outline describes the most significant seasonal changes in ground-water régime elements (level, yield, temperature, chemical composition) for comparison with long-term data, if available. Water-level fluctuation graphs, based on data for the entire period of observation, describing yearly changes in ground-water level and revealing long-term trends, might be attached. A number of hydrogeological parameters can be compiled on one diagram, one example of which is shown in Figure 7.5a.

The principal observations on the ground-water régime for a station or representative basin are presented in tabular form: (a) ground-water levels; (b) yields of flowing wells and springs; (c) ground-water temperatures; (d) ground-water chemical composition; (e) soil-moisture storage in the zone of aeration; (f) salt composition of soils in the zone of aeration; (g) ground-water balance elements.

The above data are obtained on the basis of experimental investigations. The IHD Co-ordinating Council has recommended a form of presentation of hydrological data in a series of tables under Annex XV to its report (SC/IHD/IV/36).

Distribution and grouping of factual observation data obtained at each station or basin should be as follows:

1. For unconfined ground water: according to topographical and geomorphological features such as fluvioglacial plains, alluvial terraces, piedmont plains, intermontane depressions, alluvial fans, deltas, etc. and their elements (divides, slopes, the upper part of an alluvial fan, its peripheral part, the first terrace, the second terrace, etc.).
2. For confined ground water: by the separation of aquifers according to stratigraphical subdivision, with subsequent distribution of the data according to head relationships.

The summary of factual material for each station or basin is completed by a presentation of data obtained from field or laboratory tests made during the year. Typical data would include the results of grain-size analyses, laboratory and field determinations of permeability, and data on recharging water into pits or wells.

Presentation, in the manner described above, of data on the ground-water régime comprises the contents of a summary hydrogeological yearbook of a country.

8.4.2 *Summary of data on the régime, balance and total resources of ground water*

Hydrogeological investigations which are being carried out at IHD stations provide: (a) data on the hydrogeological conditions that influence the formation of various types of ground waters in the zone of active water exchange; (b) data on the ground-water régime and balance elements and their determining factors; (c) interpretative results obtained by analysis of observation data, including the calculation of the water balance, the assessment of total ground-water resources, the prediction of the ground-water régime and the establishment of the interrelationship between ground and surface waters.

The first and second groups of data should be prepared for publication at the end of the first or the second year of investigation as the first yearbook of a hydrogeological station (Sect. 8.4.1).

The third group of data is obtained by analysis of observation data which were carried out within a period of not less than 3-5 years. These results should be prepared for publication as monographs pertaining to a certain region or as summary reports for the country as a whole.

Data from similar hydrogeological stations should be published together as a summary monograph for a larger region, which can have a number of small or medium river basins. Such a monograph should include information on the following: (a) natural conditions influencing the formation, régime and balance of ground water; (b) factors determining seasonal, yearly and long-term values and variations of the ground-water régime and balance elements; (c) the subdivision of the basin or region according to conditions influencing the formation, régime, and balance of ground waters; (d) results of analysis of the ground-water régime and calculation of the ground-water balance; (e) assessment of genetic and quantitative relations of the ground-water régime and balance elements with the surface-water régime and other natural and artificial factors; (f) development of ground-water resources; (g) forecast of future natural changes in the ground-water régime; for those areas which are affected by man's activities a forecast of the disturbed ground-water régime is made (h) specification of the future hydrogeological problems, methods for their solution and long-term research programmes.

Complex maps (see Section 8.2), hydrogeological profiles, graphs showing fluctuations of ground-water régime elements, changing values of various ground-water factors, surface-water run-off régime, etc. are included as supplements to the monograph. The publication should relate to separate aspects of the total hydrologic cycle and water balance in a basin.

Description of the natural conditions should begin with the geographic location and extent of the basin. Description of hydrography and climate should be as concise as possible.

1. Geomorphology, geological structure and hydrogeological conditions are the main subjects to be described in the first chapter. Genetic types of soils, rocks, and terrains in which ground waters occur are covered briefly in the description of these conditions.
2. The chapter of the monograph that considers the factors determining the ground-water régime must contain results obtained by the synthesis of hydrometeorological and diverse artificial factors. It is desirable to illustrate these results by means of maps or graphs that show their respective probability.
3. The chapter on subdivision of the basin or region according to ground-water-forming conditions should be based on the classification of régime factors and their influence upon the level, discharge and chemical composition of ground waters. Any subdivision of the ground-water régime based on genetic relations of the régime elements with the main influencing factors (precipitation, evaporation, drainage, etc.) should be discussed in this chapter.
4. The chapter that presents the results of analysis of the ground-water régime contains data synthesis obtained by means of application of the following methods: total water balance (Sect. 5.2); hydrological (Sect. 5.4); hydrodynamical (Sect. 6.2); analogue simulation (Sect. 6.3) and statistical (Sect. 8.4.1). The final part of the chapter is devoted to compilation of a balance for ground waters of various types in the areas of the basin and for the basin as a whole. Variations in water-balance elements and their probability of recurrence are also discussed in this chapter.
5. The chapter on the relations of the ground-water régime and balance elements with the determining factors is based on correlation of the water-régime elements with such indices as depth to water, precipitation, soil evaporation, water reserves in snow cover, water supply for irrigation, drainage modulus, etc. These relations form the basis for extrapolation of observation results over an area and for prediction of the water régime.
6. The chapter on ground-water resources assessment considers the results of a genetic separation of river hydrographs (Sect. 5.4), results of calculation of local replenishment of ground-water outflow by means of the hydrodynamic method (Sect. 5.3), etc. It is important to ascertain average long-term, maximum and minimum yearly values of ground-water recharge and storage change for years of widely varying precipitation.
7. The chapter considering forecasts of the ground-water régime is compiled taking into account the local economic requirements and the number of observations. The statistical, hydrodynamic and water-balance methods (Sect. 8.3) are widely applied for this purpose. The modelling method (Sect. 6.3) has useful application to sites that have been studied in detail. Forecasts of the ground-water régime may be made, as, for example, long-term trends in lowering and recovery of ground-water levels; long-term maxima and minima of ground-water levels; seasonal forecast of extreme ground-water levels and ground-water replenishment.

   Under conditions of irrigation or drainage, ground-water level prediction is made by means of the hydrodynamic or modelling methods depending on the planned water-management measures. In this case such a forecast is an integral part of designing a reclamative system.
8. The concluding part of a monograph contains a plan of the investigations to be carried out in the basin, with special reference to specifications and changes of those techniques which follow from the results of observations.

The most valuable results of observations should be published as scientific articles, because of their importance in the period between issues of monographs.

8.5 Computer processing of ground-water data[1]

8.5.1 *Introduction*

The purpose of data-based management systems is the collection, maintenance and manipulation of data, *en masse*. These systems provide a report-generation capability, a data-base loading facility, a teleprocessing interface, and/or a query language. There is a wide variety of existing and operational data-processing systems; consequently, selecting the system most appropriate to the needs of a national ground-water data-base management centre may entail many months of preparation and testing. Obviously, the requirements of individual countries will vary; therefore the most suitable system must be sought for each country. The following paragraphs describe the approach used by a large developed country in identifying its own national requirements.

8.5.2 *Definition of requirements*

As a first step it is necessary to identify the mandatory requirements, with which any proposed system must comply. These requirements, of course, differ in their details from one country to another. In one developed country, for example, the following mandatory requirements were identified in the selection of a national ground-water data-base management system: (a) the data-processing system had to be computer based; (b) the system chosen had to be operational and available for immediate use; (c) the system had to entail a minimum of search and updating costs; (d) proprietary data elements had to be protected; (e) the data-processing system had to be compatible with an IBM 370/155 computer in an operating system (OS); (f) the system had to be capable of storing a text with a minimum restraint on length; (g) capability for handling records of variable length in fields of variable length had to exist; (h) the records had to be selectable for any data element (inverted structure); (i) the records had to be selectable using Boolean logic; (j) the system had to handle encoding and decoding; (k) the data had to be compatible with programmes written in FORTRAN or PL/1 language; (l) capability for readily updating records had to exist; (m) the system had to be capable of adding new data elements without re-creating the entire data base; (n) capability for retrieving records in sorted order had to exist; (o) the system had to have adequate file management and maintenance procedures; (p) the system had to be optimized so as to minimize I/O (input/output) operations.

8.5.3 *Scope of requirements*

As a second step it is necessary to identify as many software packages as possible that fit the defined requirements for data handling. In the developed country previously mentioned seventy-six data-processing systems were found to exist. However, a detailed review of these seventy-six systems indicated that only eight of these systems were available for data-base management. After eliminating those systems that were not available on the IBM 360 or 370 series of computers and those that did not provide an interface capability with the host language, the list was reduced to six. The remaining six

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[1] This section has been prepared by C. R. Showen and C. M. Baker and revised by a panel of editors composed of: A. A. Konoplyantsev, E. P. O'Driscoll, H. J. Schoeller and G. C. Taylor.

systems were further compared using the following criteria: M, mandatory, a 'must-have' feature; D, desirable, not mandatory but a feature that should be available; L, 'like-to-have', would be nice to have but immediate requirements could be implemented without; C, characteristic for which special consideration is given during the evaluation; not wholly necessary.

8.5.4 *Individual systems evaluation*

Using the above criteria two systems were selected from the six that most nearly satisfied the requirements. These two systems were ADABAS and SYSTEM 2000. ADABAS directly satisfied all sixteen mandatory (M) requirements. It also satisfied eleven desirable (D) features. The four desirable features that were not satisfied were (a) auto encoding; (b) remote batch (retrieval); (c) auto decoding; and (d) record security. ADABAS also had two of the seven like-to-have (L) features. SYSTEM 2000 directly satisfied fourteen of the sixteen mandatory requirements. The requirement for capability in handling variable length records was not satisfied. SYSTEM 2000 satisfied eleven desirable features; those not satisfied being (a) data flexibility; (b) data extensibility; and (c) data loading facility. The system also had four of the seven like-to-have features.

8.5.5 *Benchmark testing*

Based on the foregoing evaluations the ABADAS and SYSTEM 2000 were chosen for benchmark testing on an IBM 155 computer at the national centre. The benchmark tests included the following tasks: (a) addition of new records; (b) a set of queries; (c) low-volume modification of existing data; (d) interfacing of PL/1 programmes to the data base; (e) modification of the content of the data base; (f) backups and recovery (in the event the data base is accidentally damaged); (g) reallocation of disc storage space.

Based on the results of these benchmark tests, the study team recommended that the SYSTEM 2000 be selected for automatic data processing (ADP), storage and retrieval of ground-water data throughout the country.

8.5.6 *National ground-water site-inventory data base*

8.5.6.1 **Data-base description**

SYSTEM 2000 is a generalized data-base management system which is used to manage and maintain a national ground-water site-inventory data base. This base contains physical, hydrological and geological data about sites where ground water is, or can be, withdrawn from an aquifer, or sites where potential aquifers are exposed at the surface. Most such sites are wells, but test holes, springs, tunnels, drains, ponds may also exist. Other excavations and outcrops are also included, in addition to specialized installations such as well batteries and collector wells. The definition of the data base is shown in Table 8.5.6.

8.5.6.2 **Data-base structure**

The logical organization of a SYSTEM 2000 data base is a hierarchical structure with inverted lists or indices. All the data about a single major item (in this case, a well, a spring, or other ground-water site) comprise a logical entry in the data base. Individual pieces of information, such as the altitude of the site or the temperature of the water, are data elements; logically related data elements are grouped into blocks called repeating groups.

Under the concept of the hierarchical structure every data element belongs to a repeating group, and every repeating group belongs to a higher level repeating group (the logical entry is, itself, a repeating group at the highest level). For example, information about the casings in a well belongs to the set of information about the well's construction; the construction information, in turn, belongs to the well. The ground-water data base has three levels of repeating groups, which are indicated in Table 8.5.6 by three degrees of indentation. In the data-base definition given in Table 8.5.6, repeating group names are indicated by the letters (RG), and the data elements under them are indented. A fundamental aspect of the repeating group is that it can occur once, many times or not at all. Each logical entry contains all the information about an individual site, and because it is a repeating group itself, there can be information about many sites in the file.

8.5.6.3 Data retrieval

To facilitate retrieval of the data, inverted lists or indices are constructed by the system, based on the values of key elements indicated in Table 8.5.6 by (KEY). A table is constructed for each key element listing each unique value for that element, the number of logical entries which contain that value, and the direct-access address of those logical entries in the main data storage area. Some key elements have a unique value for each entry and only one entry per unique value; for example, SITE-ID. (Records with duplicate SIDE-IDs will not be stored in the data base.) Other key elements, such as the code for DISTRICT, will have a few unique values, but many occurrences of each value.

This indexing permits rapid retrieval of selected logical entries, with minimum access to the main body of the data. Certain statistical information about key elements, such as the number of logical entries that contain a given value or the range of unique values for a given key element, are determined from the inverted lists, and can be obtained without access to the main body of data.

An important aspect of the system is that logical entries are of varying length, because if none of the data elements in a repeating group has a value, the repeating group is not stored and no space is reserved for it. If information belonging to a non-existent repeating group is added later, the repeating group is stored with a pointer to relate it to the correct logical entry.

8.5.6.4 Data-element formats

The field length for numeric values and dates is fixed, but for textual data the system allows for entries longer than the specified field length. When a string of characters longer than the specified field size is entered, excess characters are stored elsewhere in the data base and a pointer to the extra characters is stored in place of the value. This feature, called overflow, can be valuable for occasional use, but over-use of the feature degrades the performance of the system because it requires extra accesses to the data base to retrieve the overflow values. The maximum length permitted for any field is eighty characters.

8.5.6.5 Data input

All input to the data base is made through an input-edit programme that checks the data for internal consistency. Eleven data elements (preceded by an asterisk in Table

TABLE 8.5.6. Ground-water site-inventory data-base definition

```
SITE-ID (KEY)
SITE-TYPE (KEY)
RECORD-CLASSIFICATION (KEY)
SOURCE-AGENCY (KEY)
PROJECT-NUMBER (KEY)
DISTRICT (KEY)
STATE
COUNTY
*STATE-COUNTY (KEY)
LATITUDE (KEY)
LONGITUDE (KEY)
COORDINATE-ACCURACY
LOCAL-NUMBER (KEY)
LAND-NET-LOCATION
LOCATION-MAP-ID
LOCATION-MAP-SCALE
ALTITUDE (KEY)
ALTITUDE-METHOD
ALTITUDE-ACCURACY
TOPOGRAPHIC-SETTING
OWDC-HYDROLOG-UNIT (KEY)
DATE-CONSTRUCTED (KEY)
*DATE-CONST-ACC
SITE-USE
WATER-USE (KEY)
SECOND-WATER-USE
THIRD-WATER-USE
  OPENINGS (RG)
    OPENING-TOP (KEY)
    OPENING-BOTTOM (KEY)
    OPENING-TYPE (KEY)
    SCREEN-MATERIAL
    OPEN-DIAMETER
    OPENING-WIDTH
    OPENING-LENGTH
GEOHYDROLOGIC-UNITS (RG):
  GEOHYDRO-TOP (KEY)
  GEOHYDRO-BOTTOM
  GEOHYDRO-UNIT (KEY)
  LITHOLOGY (KEY)
  LITH-MODIFIER
  AQUIFER (RG)
    AQUIFER-DATE
    AQUIFER-DATE-ACCURACY
    AQUIFER-STATIC-LEVEL
    AQUIFER-CONTRIBUTION (KEY)
HYDRAULIC-DATA (RG):
  HYDRAULIC-SEQ-NO (KEY)
  HYDRAULIC-UNIT-ID (KEY)
  TEST-INTERVAL-TOP
  TEST-INTERVAL-BOTTOM
  HYDRAULIC-UNIT-TYPE
  HYDRAULIC-REMARKS
  COEFFICIENTS (RG)
    COEFFICIENTS-SEQ-NO
    TRANSMISSIVITY
    HORIZONTAL-CONDUCTIVITY
    VERTICAL-CONDUCTIVITY
    STORAGE-COEFFICIENT
    LEAKANCE
    DIFFUSIVITY
    SPECIFIC-STORAGE
QUALITY-NETWORK (RG):
  QN-BEGIN-YEAR
  QN-END-YEAR
  QN-DATA-SOURCE
  QN-FREQUENCY
  QN-ANALYSIS-TYPE (KEY)
LEVEL-NETWORK (RG):
  LN-BEGIN-YEAR
  LN-END-YEAR
  LN-DATA-SOURCE
  LN-FREQUENCY
PUMPAGE-NETWORK (RG):
  PN-BEGIN-YEAR

HOLE-DEPTH (KEY)
WELL-DEPTH (KEY)
WELL-DEPTH-SOURCE
WATER-LEVEL
WATER-LEVEL-DATE
*WL-DATE-ACCURACY
WATER-LEVEL-SOURCE
WATER-LEVEL-METHOD
SITE-LEVEL-STATUS
PUMP
GEOHYDRO-DATA-SOURCE
*LAST-UPDATE
*VERIFIED (KEY)
LIFT (RG):
  LIFT-TYPE
  LIFT-DATE
  *LIFT-DATE-ACCURACY
  INTAKE-SETTING
  POWER-TYPE (KEY)
  HORSEPOWER (KEY)
  MAJOR-PUMP (RG)
    MANUFACTURER
    SERIAL-NUMBER
    POWER-COMPANY
    ACCOUNT
    METER
    CONSUMPTION
  PN-END-YEAR
  PN-DATA-SOURCE
  PN-FREQUENCY
  PN-DATA-METHOD
FLOW-DATA (RG):
  FLOW-SEQ-NO
  FLOW-MEAS-DATE
  *FLOW-DATE-ACC
  FLOW-DISCHARGE (KEY)
  FLOW-DISCHARGE-SOURCE
  FLOW-DISCHARGE-METHOD
  FLOW-PROD-LEVEL
  FLOW-STATIC-LEVEL
  FLOW-LEVEL-SOURCE
  FLOW-LEVEL-METHOD
  FLOW-PERIOD
PUMP-PRODUCTION (RG):
  PUMP-SEQ-NO
  PUMP-MEAS-DATE
  *PUMP-DATE-ACC
  PUMP-DISCHARGE (KEY)
  PUMP-DISCHARGE-SOURCE
  PUMP-DISCHARGE-METHOD
  PUMP-PROD-LEVEL
  PUMP-STATIC-LEVEL
  PUMP-LEVEL-SOURCE
  PUMP-LEVEL-METHOD
  PUMP-PERIOD
OWNERS (RG):
  OWNERSHIP-DATE
  *OWNERSHIP-DATE-ACCURACY
  LAST-NAME
  FIRST-NAME
  MIDDLE-INITIAL
MINOR-REPAIRS (RG):
  REPAIR-SEQ-NO
  REPAIR-NATURE (KEY)
  REPAIR-DATE
  *REPAIR-DATE-ACCURACY
  REPAIR-CONTRACTOR
  PERFORMANCE-CHANGES
SPRINGS (RG):
  SPRING-NAME
  SPRING-TYPE (KEY)
  PERMANENCE
  DISCHARGE-SPHERE
  IMPROVEMENTS
  NUMBER-SPRING-OPENINGS

    PUMP-MAINTAINER
  STANDBY (RG)
    STANDBY-POWER-TYPE
    STANDBY-HORSEPOWER
CONSTRUCTION (RG):
  CONST-SEQ-NO (KEY)
  DATE-COMPLETED (KEY)
  *CONST-DATE-ACCURACY
  CONTRACTOR
  CONST-DATA-SOURCE
  CONST-METHOD (KEY)
  FINISH
  SEAL-TYPE
  SEAL-BOTTOM
  DEVELOPMENT-METHOD (KEY)
  DEVELOPMENT-DURATION
  SPECIAL-TREATMENT (KEY)
  HOLES (RG)
    HOLE-TOP
    HOLE-BOTTOM
    HOLE-DIAMETER
  CASINGS (RG)
    CASING-TOP
    CASING-BOTTOM
    CASING-DIAMETER
    CASING-MATERIAL (KEY)
    CASING-THICKNESS
  FLOW-VARIABILITY
  FLOW-VAR-ACCURACY
OTHER-DATA (RG):
  TYPE-DATA
  DATA-LOCATION
REMARKS (RG):
  REMARK-DATE
  REMARK
SITE-VISITS (RG):
  INVENTORY-DATE
  INVENTORY-PERSON
OTHER-IDS (RG):
  OTHER-ID (KEY)
  OTHER-ID-ASSIGNER
FIELD-WATER-QUALITY (RG):
  FWQ-SAMPLE-DATE (KEY)
  *FWQ-DATE-ACC
  FWQ-GEOHYDRO-UNIT (KEY)
  FWQ-PARAMETER (KEY)
  FWQ-MEASUREMENT (KEY)
LOGS (RG)
  LOG-TYPE (KEY)
  LOG-TOP
  LOG-BOTTOM
  LOG-SOURCE
WELL-GROUP (RG):
  NUMBER-WELLS
  WG-DEEPEST
  WG-SHALLOWEST
  WG-METHOD
POND-TUNNEL-DRAIN (RG):
  PTD-LENGTH
  PTD-WIDTH
  PTD-DEPTH
COOPERATOR-DATA (RG):
  COOPERATORS-ID (KEY)
  CONTRACTOR-REGISTRATION-NO
  INSPECTION-STATUS (KEY)
  REASON-UNAPPROVED
  DATE-INSPECTED
  COOPERATOR-REMARKS
LATERALS (RG):
  LATERAL-NUMBER
  LATERAL-DEPTH
  LATERAL-LENGTH
  LATERAL-DIAMETER
  LATERAL-MESH
```

8.5.6), which appear in the data-base definition, do not appear on any input forms and are never initiated by the user. These fields are generated by the input-edit programme. Their values and uses are as follows:

STATE-COUNTY: This field is a retrieval key by which the state and county codes are joined together. Use of this field for retrievals on a state and county basis is much more efficient than using state codes and county codes separately. (State and county codes, accordingly, are NON-KEY.)

LAST-UPDATE: Whenever any change is made in a logical entry, including the first, this field is set to the current date, so that the user will know how up to date is the information that he retrieves.

VERIFIED: Whenever any change is made in a logical entry, this field is set to NO-data not verified. When the changed data, which is always printed out, has been visually verified by the originating office, another job must be run to set the VERIFIED field to OFF-data verified. This field provides the user with an idea of the validity of the data that he retrieves. It is very important that the data entered into the system be visually verified and the VERIFIED switch set to OFF; unverified data may be correct but they are always suspect.

The remaining eight fields, marked by an asterisk in Table 8.5.6, are all data-accuracy codes. In SYSTEM 2000, the fields for dates must always contain complete, valid dates. It often happens that data are available for water levels, well construction, discharge and the like, for which only part of the date is know such as the year only, or month and year but not the day. In order to store the data, a valid month and/or day must be supplied, which raises the question of the legal validity of the date. For these situations, the missing day and/or month are entered on the input forms as 00. The input-edit programme substitutes values of 01 for these 00 values (making a valid date) and enters a date accuracy code to reflect the accuracy of the original date. The code is set to 'D', if the input date was accurate to the day, to 'M', if the date was accurate only to the month, or to 'Y', if the date was correct only to the year. If the user retrieves these date-accuracy fields together with the dates, he has some information about the precision with which the date is known.

8.5.6.6 Data output

SYSTEM 2000 is very much oriented towards the needs of the user. Retrieval from the data base can be carried out in several different forms. Using the retrieval language built into SYSTEM 2000, data can be retrieved selectively and listed in a variety of ways. These retrievals can be run both in batch mode (retrieval request on cards, output on line printer) and in time-share mode (request and output on typewriter-type terminal). Although most data selection will be on the values of KEY fields, the retrieval logic also allows selections based on the values of NON-KEY elements. Programmes are also provided to retrieve selected data, to prepare publications tables, and to pass selected data to STATPAC and to the plotter routines.

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9 Application of geophysical techniques in ground-water investigations[1]

9.1 Geophysical prospecting

9.1.1 *Introduction*

In the 1950s, geophysical prospecting in hydrogeology was generally limited to siting locations for boreholes in small areas close to towns or industrial plants. Since then, the rapid expansion of urban and industrial needs and the development of irrigation have led to a greater exploitation of ground water which sometimes results in dangerously lowering the water level and increasing the salinity. The necessity of evaluating the extent of the aquifers and the amount of water available has considerably enlarged the role of geophysical prospecting in hydrogeology. Geophysical investigations will never replace drilling but they are relatively less expensive and can provide useful information on significant features of subsurface stratigraphy and structure. This reduces the number of exploratory boreholes and helps in predicting their depths.

In hydrogeology, geophysical prospecting can help to: locate beds or fractures below an unconformable overburden; estimate the depth of an aquifer and of the hydrostatic level; determine the lateral extent, the thickness and the volume of a formation; evaluate the total porosity and the water stored in an aquifer; estimate the water salinity; determine the connexions between aquifers; trace subterranean streams; locate submarine fresh-water springs; locate the origin of thermomineral waters.

The design of a geophysical survey should be prepared by a hydrogeologist and a geophysicist working together. The hydrogeologist sets out the purposes of the survey and supplies the geophysicist with all the available geological and hydrological data. The geophysicist then decides whether geophysics can help and if so chooses the best method or combination of methods and prepares a field programme.

In general, some boreholes are necessary to calibrate the geophysical results. Existing boreholes can be used, but if there are none, a limited drilling programme should be planned to help the interpretation.

The electrical resistivity method and the seismic refraction methods are the most important in hydrogeology; the other methods used are: seismic reflection, gravity prospecting, ground thermometry and airborne infra-red thermometry, airborne radio-wave methods and the magneto-telluric method.

The inductive methods (alternating current-input) and the induced polarization method have not been included because they were developed primarily for mining

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[1] This chapter, apart from the section on nuclear logging prepared by V. I. Ferronsky, Director of Reseach and Laboratories, International Atomic Energy Agency, Vienna (Austria), was written by J. L. Astier. It was revised by a panel of editors composed of A. A. Konoplyantsev, E. P. O'Driscoll, H. J. Schoeller and G. C. Taylor. Many of the illustrations have been reproduced from J. L. Astier, *Géophysique Appliquée à l'Hydrogéologie*, Paris, Masson 1971.

surveys. Some attempts have been made to apply them to hydrogeology but without practical results.

It is impossible to compare significant costs of the different geophysical operations because salaries, equipment, field production and the quality of the field measurements and of the interpretation vary too much.

9.1.2 *Ground geophysical prospecting*

In general, results of two or more independent geophysical techniques are often used, the best combination being dependent on local conditions. However, some suggestions are provided where appropriate.

9.1.2.1 **Electrical methods**

Resistivity method. Resistivity of rocks. The immediate purpose of the resistivity method is to determine the distribution of resistivities in various rock types of the earth.

Table 9.1.2.1 gives the usual limits for the resistivity of some types of water and of the more common rocks. The resistivity of rocks varies from less than one to several thousand ohm m, which helps to distinguish between different types of rocks.

Except for graphitic schists, the conductivity of rocks is of electrolytic origin; it increases with the total porosity and the salt content of the contained water. The manner in which the water is distributed also influences the conductivity. Archie (1942) has experimentally shown the two following relationships between ρ, the resistivity of a non-argillaceous aquifer, ρ_w, the ground-water resistivity, and θ, the total porosity: $\rho = F\rho_w$ and $F = 1/\theta^n$, where F is called the 'formation factor' and n the 'pore distribution coefficient'.

Combining the two equations, one finds: $\theta^n = \rho_w/\rho$.

Figure 9.1.2.1a gives typical curves obtained from Archie's formula, using values for the coefficient n.

The resistivity of stratified rocks also depends on the direction of flow of the electric current. There may be anisotropy due to the microstructure of the rocks or caused by thin sedimentary beds if these are alternately resistive and conductive. The lowest and highest values correspond to the current flowing respectively parallel and perpendicular to the bedding planes. The ratio between these two extreme values sometimes reaches several units.

TABLE 9.1.2.1. Resistivity of waters and rocks

Waters or rocks	Resistivity (ohm m)	Waters or rocks	Resistivity (ohm m)
Sea water	0.2	Clayish sandstones	50–300
Spring water	50–100	Sandstones, quartzites	300–10,000
Dry sands and gravels	1,000–10,000	Volcanic ashes, tuffs	20–100
Sands and gravels under fresh water table	50–500	Lavas	300–10,000
		Graphitic schists	0.5–5
Sands and gravels under saline water table	0.5–5	Clayish or weathered schists	100–300
		Schists	300–3,000
Clay	2–20	Weathered gneiss or granites	100–1,000
Marls	20–100	Gneiss, granites	1,000–10,000
Limestones	300–10,000		

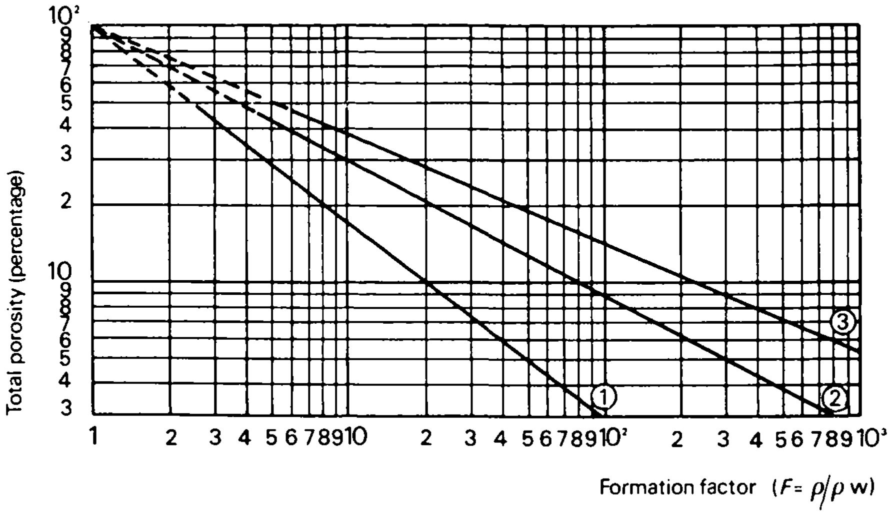


FIG. 9.1.2.1.a. Curves giving the total proposity of a saturated non-argillaceous aquiferous formation as a function of the formation factor F: curve 1, loose formation; curve 2, consolidated rock with interstitial porosity; curve 3, consolidated rock with fissure porosity.

Resistivity measurement. The ground resistivity is measured using a quadripole AMNB configuration (Fig. 9.1.2.1b). A direct current or a very low-frequency alternating current, of intensity i, is passed through the ground by electrodes A and B and the resulting potential difference ΔU between electrodes M and N is measured. The ground resistivity ρ is given by the following formula:

$$\rho = K \frac{\Delta U}{i}$$

where K is a geometric coefficient depending only upon the relative position of the four electrodes.

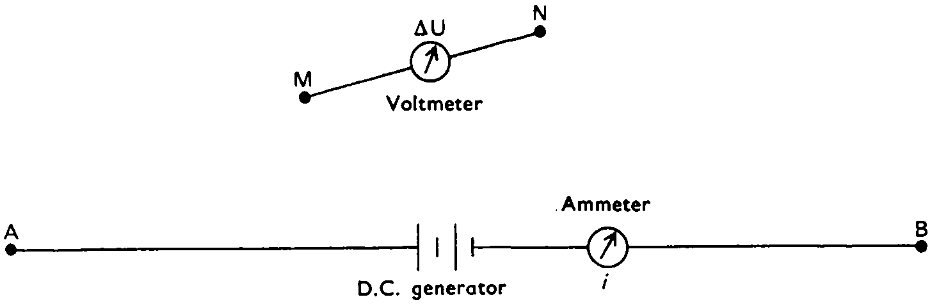


FIG. 9.1.2.1b. Quadripole AMNB configuration for resistivity measurements.

The four electrodes are usually laid out using the Schlumberger configuration. The four electrodes are in line, the centres of AB and MN coincide and MN is always small with respect to AB (less than a third). For this configuration $K = \pi\, AM \,.\, AN/MN$. The Wenner configuration, in which the ratio AB/MN is constant and equal to 3, is now less frequently used because the graphs are much more disturbed by the soil variations below M and N than are those obtained with the Schlumberger configuration. The dipole array is only used for very deep investigations, when for practical reasons one wants to avoid laying out a very long line AB.

Direct current is commonly preferred because alternating current presents the major disadvantage of a depth of penetration which decreases rapidly with the frequency.

The resistivity obtained corresponds to the true resistivity only if the ground is homogeneous. When the ground is not homogeneous, which is obviously the usual case, the measured resistivity is an apparent resistivity ρ_a related to the true resistivity and thickness of all the beds which influence the current flowing between A and B. The apparent resistivity ρ_a is sometimes defined as the true resistivity of an imaginary homogeneous and isotropic material equivalent to the real heterogeneous one.

The study of the variations in apparent resistivity is the basis of the resistivity methods. Two techniques of sounding and profiling are employed depending upon whether vertical or horizontal information is desired.

Electrical sounding. Electrical soundings are used to study the variation of resistivity with depth. The depth of penetration is controlled by the separation of the current electrodes AB which carry the current. Figure 9.1.2.1c illustrates how the paths of current penetrate deeper and deeper beds as the electrode separation is increased.

An electrical sounding consists of a succession of apparent resistivity measurements made with an increasing separation length AB while the centre of the quadripole and its azimuth remain fixed. On log-log paper, a graph is established by plotting the apparent resistivity ρ_a against a constant fraction of the length AB, usually AB/2.

The choice of AB/2 for the abcissa is arbitrary and it does not mean that the depth of investigation is equal to AB/2. In fact, the depth of investigation varies in a complex way according to the number of strata involved and the ratios of their resistivity and thicknesses. In most cases, it falls between AB/4 and AB/12.

The use of logarithmic co-ordinates for plotting is justified by the following reasons: (a) the effect of a bed decreases with depth; (b) the variations in resistivity of conductive and resistive beds are equally important; (c) the master curves calculated for a geological model can be used to interpret all the curves corresponding to geological sequences for which the ratios of the thicknesses or resistivities are identical.

Figure 9.1.2.1d gives three electrical soundings showing the apparent dip of a bed which has a constant thickness.

The quantitative interpretation of the electrical soundings is made by matching the field graphs with master curves. In this way the geophysicist obtains a value of the resistivity ρ and the thickness m of the uppermost layer, the transversal resistance (or product e times ρ) of the next resistive layers and the longitudinal conductance (or ratio m/ρ) of the next conductive layers. A layer appears on the electrical sounding only if it is thick enough in respect to its depth.

On the other hand, the interpretation is facilitated by the following conditions: (a) if a deep layer is very thick, its resistivity and thickness are obtained directly from the master curves; (b) field graphs give directly the total longitudinal conductance of the beds overlying a resistive substratum; (c) electrical soundings corresponding to geological series with four or more layers can be interpreted with master curves calculated for two and three layer models. The process is taken step by step, the shallowest layers, the thickness and resistivity of which have already been calculated, being combined and treated is a single equivalent layer.

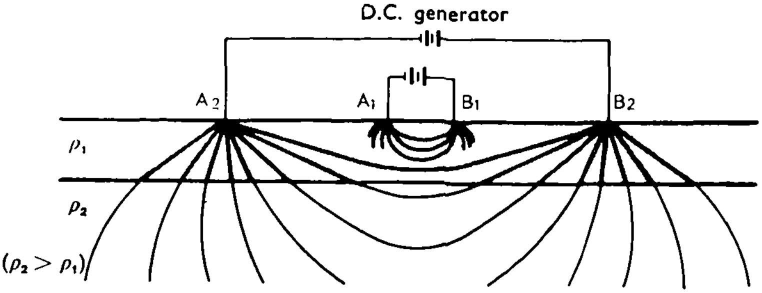


FIG. 9.1.2.1c. The depth of investigation increases with the distance AB.

Computers are sometimes used to verify whether the proposed interpretation consistently matches the field measurements. Automatic processing can be used to give out physical solutions, and the geophysicist has to adapt these to the geological constraints.

Resistivity profiling. A resistivity profile consists of a series of resistivity measurements made with a quadripole AMNB configuration of a constant length but moved along a straight line. The distance between each measurement is constant. The direction of line AB is the same as that of the profile. Between two observations the line MN is generally moved a distance equal to MN in order to have adjacent measurements and to realize a continuous investigation. With several profiles it is possible to draw an apparent resistivity map by drawing contours through the points of equal value.

The resistivity profiles and maps can represent the result of a horizontal investigation at a constant depth. They are particularly useful for finding geological boundaries hidden below overburden and for indicating steeply inclined contacts and faults.

Figure 9.1.2.1e shows the resistivity profile obtained in a laboratory by using a small-scale model constructed to represent a simple faulted structure.

Tulúm Valley (Argentina). The Tulúm Valley of the San Juan Province is one of five pilot areas studied by Plan Agua Subterranea, a joint project of the Republic of Argentina

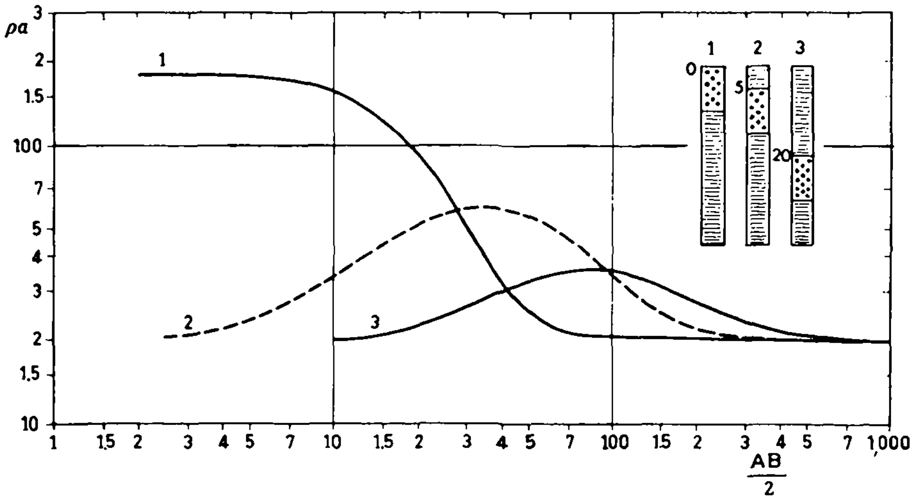


FIG. 9.1.2.1d. Electrical soundings, showing the dip of a bed with constant thickness.

and the United Nations Special Fund. In order to construct a mathematical model for future management of the valley, the geometry and the volume of the water-bearing materials were needed. These data were obtained by using a resistivity survey with calibration boreholes (electric logging of the boreholes was done). About 200 electrical soundings were carried out on a line AB reaching sometimes 6,000 m.

Figure 9.1.2.1f is an interpreted cross-section. This section crosses a fault which separates two basins. In the western basin only two wells have tapped the deep and thick tertiary sands and gravels characterized by a resistivity of 300–500 ohm m. In the eastern basin, the main reservoir is at a lower depth, and it has been cut by several boreholes. Its resistivity is only 100 ohm m due to the presence of fine material.

Figure 9.1.2.1g is a map giving the main structural features of the valley, the depth of the base of the reservoir and the quality of the water-bearing formation (the higher the resistivity of this formation, the lower its clay content).

Marismas area (Spain). The Marismas area is the delta of the Guadalquiver river. It was studied for the development of the Guadalquivir basin in the joint project of the Spanish Government and the Food and Agriculture Organization of the United Nations.

Sands and gravels about 100 m thick contain water with a total dissolved salt content (TDS) varying between less than 0.5 g/l to more than 10 g/l. Profiles of electrical soundings along the line AB reaching 2,000 m were carried out to delimit the area where the TDS is less than 1.5 g/l.

The position of the fresh/saline water interface was worked out using the graph of Figure 9.1.2.1h. On this figure the TDS has been plotted against the aquifer's transverse resistance for ten boreholes. The ten points obtained are close to a line corresponding to an aquifer 100 m thick and with a total porosity of 20 per cent. The absolute error on the TDS determined with the theoretical line is less than 0.3 g/l except for borehole 1018 44. For this last borehole, the difference reaches 0.6 g/l, but that can be due to the fact that the transverse resistance used has been calculated from an electrical sounding situated at 500 m from the borehole.

The isoresistance lines 2,300, 3,000 and 4,600 ohm m^2 were drawn from the electrical soundings, and correspond to a TDS of 2, 1.5 and 1 g/l.

Yamaoussoukro (Ivory Coast). The village of Yamaoussoukro is situated in a granitic zone. Using a base line AB = 400 m a resistivity map has been drawn (Fig. 9.1.2.1i) which shows a main fault F1 separating an area with an apparent resistivity of 1,000 to 4,000 ohm m from an area with an apparent resistivity of 500 to 1,500 ohm m. A borehole has been drilled on each side of the fault. Borehole S1, drilled in the conductive area, encountered a relatively great thickness of weathered granite and only reached the compact rock at 56 m. It yields 6 m^3/h which is a good result for the zone. Borehole S2 drilled in the resistive area reached the bedrock at a shallow depth and was practically dry.

Magneto-telluric method. The magneto-telluric method consists of determining the apparent resistivity ρ_a of the ground using the telluric field and the associated magnetic field. The apparent resistivity is given by the following formula (Cagniard, 1953):

$$\rho_a = 0.2\, t \left(\frac{E_x}{H_y}\right)^2$$

where:

t = period of the telluric current;

E_x = the component of the telluric field for the direction Ox;

H_y = the component of the magnetic field for the direction Oy, at right angles to Ox.

The depth of investigation H_t is controlled by the time period of the telluric currents

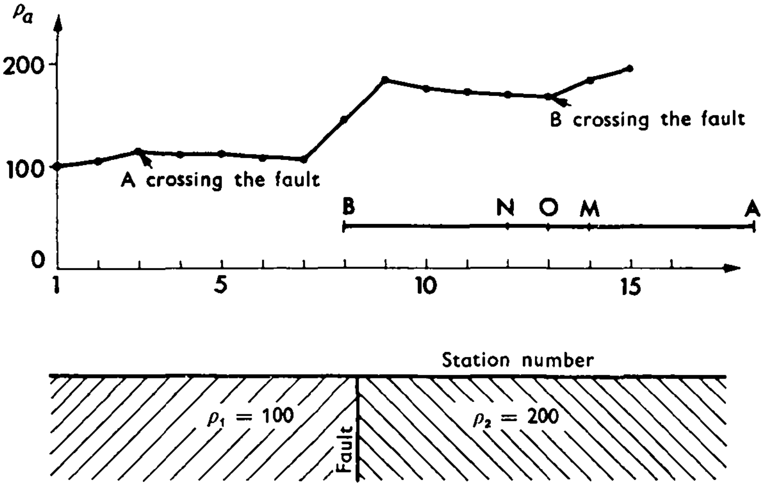


FIG. 9.1.2.1e. Resistivity profile over a vertical fault. (Compagnie Générale de Géophysique.)

considered: the longer the period, the greater the depth of investigation. In first approximation this depth is equal to

$$\frac{1}{2\pi}\sqrt{10\rho t}.$$

As with resistivity, both magneto-telluric soundings and magneto-telluric profiles are possible. Magneto-telluric soundings give the variation of the apparent resistivity at one point in relation to the period t or $\sqrt{t}$. Magneto-telluric profiles consist of measurements made with a constant time period at regular intervals along a line.

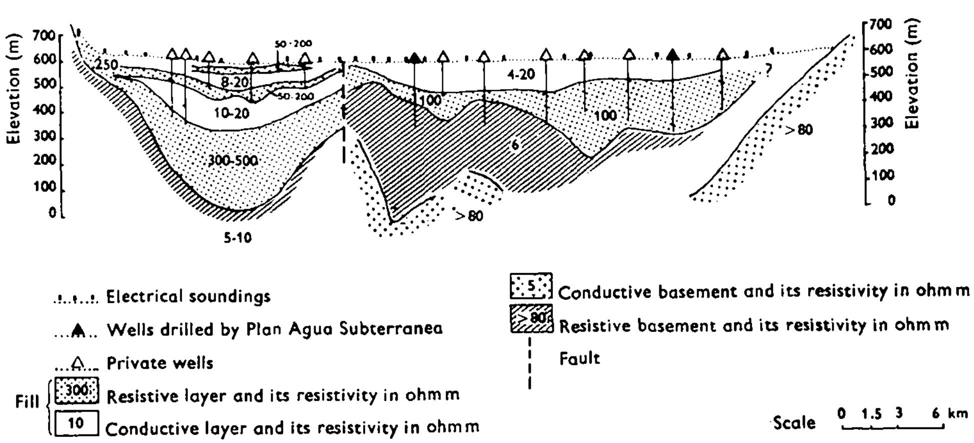


FIG. 9.1.2.1f. Tulúm Valley: interpretation cross-section. (European Association of Exploration Geophysicists.)

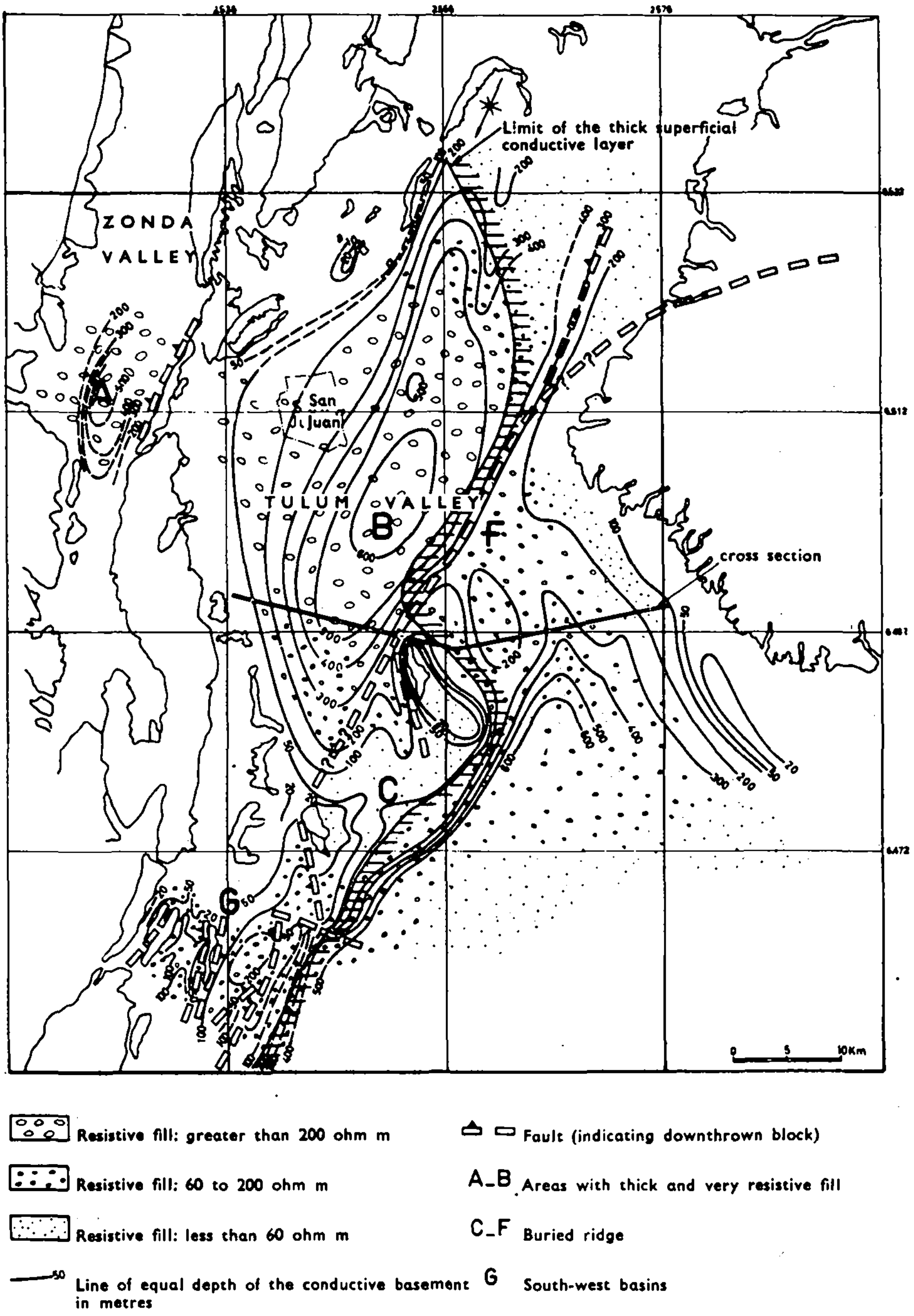


FIG. 9.1.2.1g. Tulúm Valley: map of the resistive fill and of the conductive basement. (European Association of Exploration Geophysics.)

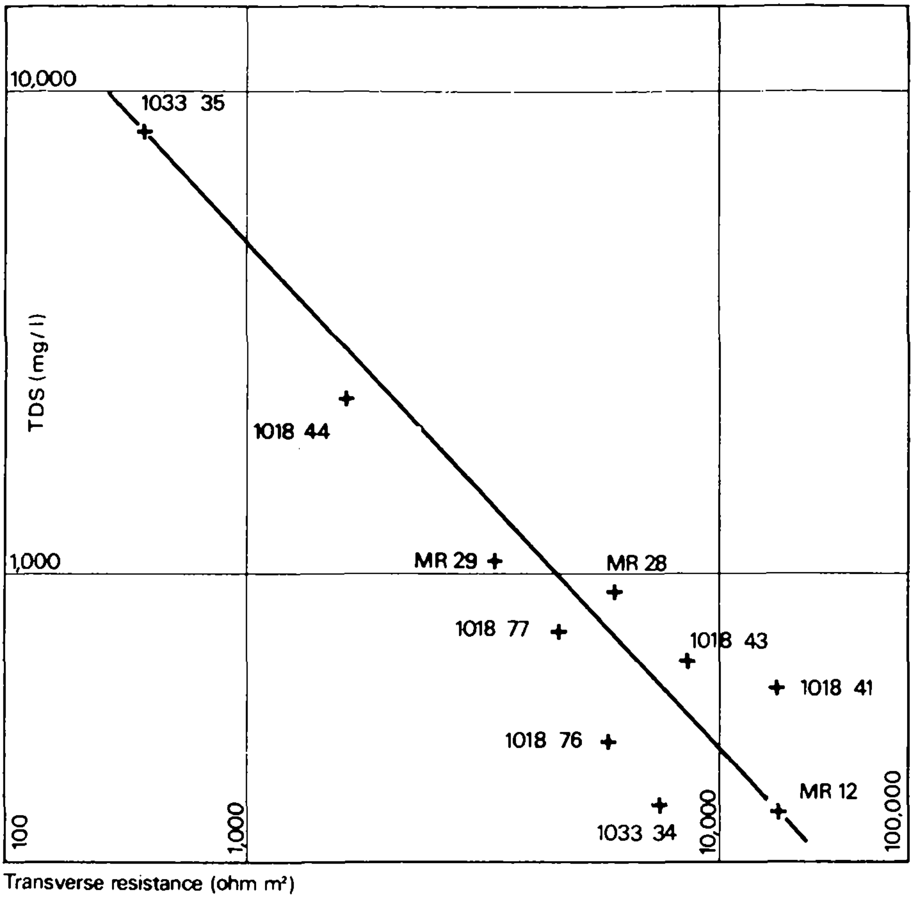


FIG. 9.1.2.1h. Marismas. Curve giving the TDS as a function of the transversal resistance of the water-bearing formation: theoretical curve—aquiferous formation, 100 m thick with a total porosity of 20 per cent. + = boreholes. (FAO.)

Magneto-telluric soundings are interpreted using theoretical master curves. Conductive layers have an effect mainly because of their longitudinal conductance *m/ρ* and the resistive ones mainly because of their thickness.

The measurement of the magnetic field is much more difficult for short periods than for long ones. For this reason the magneto-telluric method has been used mainly by petroleum companies for very deep investigations. Furthermore, the main advantage of the magneto-telluric method over the resistivity method is that it does not need a long current-carrying cable, a great advantage for deep exploration.

In the last few years, interesting methods have been developed using an artificial field instead of the telluric field but no complete field surveys have been published.

9.1.2.2 Seismic prospecting

Principle. If the ground is vibrated at one point, elastic waves are created. The time necessary for the disturbance to reach a receiver depends upon the nature and structure

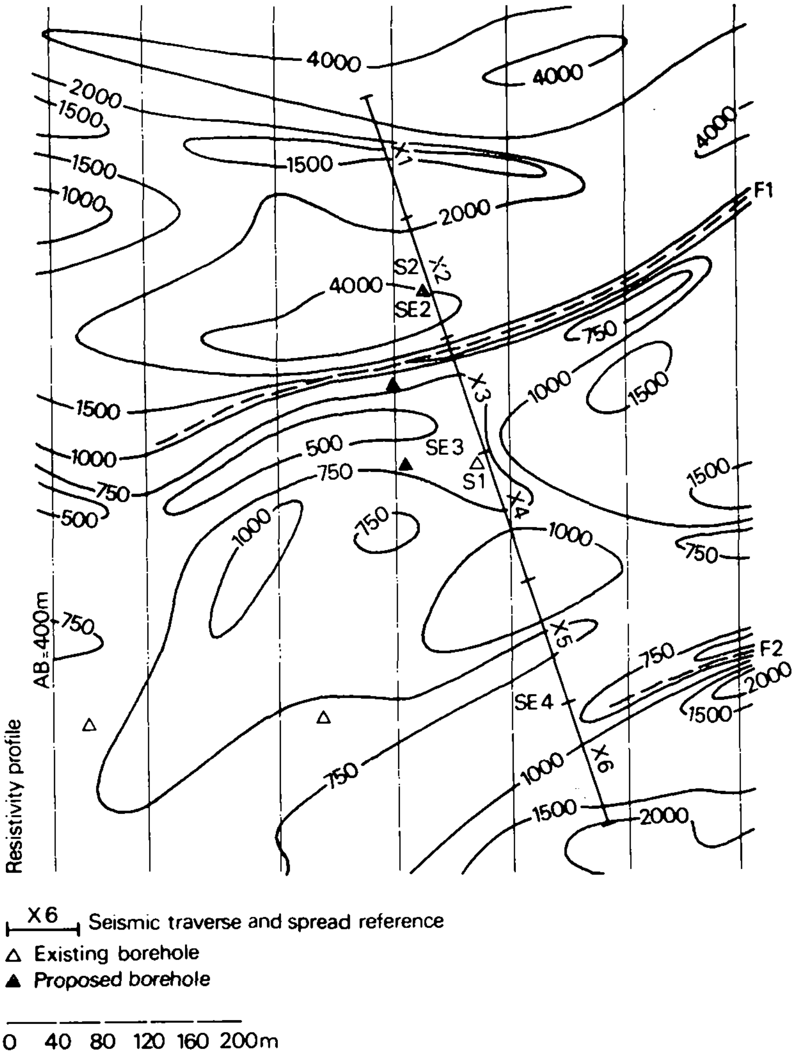


FIG. 9.1.2.1i. Yamaoussoukro: apparent resistivity map with AB = 400 m. (Comité Interafricain d'Etudes Hydrauliques.)

of the rocks. Seismic prospecting is based on the analysis of this propagation time.

Among the elastic waves, only the compressional or longitudinal waves are considered as these are the fastest.

At the interface of two media with different velocities, elastic waves are reflected and refracted in the same way as light waves and obey the following laws: (a) the reflected and refracted raypaths are in the same plane as the incident raypath and the perpendicular to the interface at the incident point; (b) the reflected angle equals the incident angle; (c) the refracted angle i_2 is related to the incident angle i_1 by the following formula: $\sin i_1/\sin i_2 = V_1/V_2$, V_1 and V_2 being the velocities of the two media. When i_1 is equal to the critical angle, $\sin i_c = V_1/V_2$, and the refracted wave propagates along the boundary ($i_2 = 90°$).

The ground vibration is usually created by a small explosion at shallow depth. Sometimes a hammer or an impacter is used. The detectors, or geophones, are electromagnetic devices transducing the earth's vibrations into an alternating current which is amplified and then sent to a recorder. The time it takes for the disturbance to arrive at the different geophones is recorded with the explosion time, known as time break, on paper with reference lines which have a spacing of 10 milliseconds.

The velocity of the elastic waves in the ground ranges between 0.3 and 6 km/sec. For a particular type of rock it decreases with the degree of fracturing or alteration and tends to increase with depth and geologic age. It is higher beneath the water table than above it, and within saturated rocks the velocity decreases with increased porosity. Table 9.1.2.2 gives velocities for the more common earth materials.

TABLE 9.1.2.2. Elastic-wave velocity

Material	Velocities (km/sec)	Material	Velocities (km/sec)
Air	0.33	Marls	2–3
Water	1.45	Sandstones	2–3.5
Weathered zone	0.33–0.6	Lavas	2.5–4
Dry alluvial deposits	0.6–1.2	Limestones and dolomites	3–5
Saturated alluvial deposits	1.6–2.4	Schists, micaschists	3–4.5
Clays	1.8–2.4	Gneiss, quartzites	3–5.5
Volcanic tuffs	1.8–2.5	Granite	4–6

Seismic prospecting can use either of two methods according to the waves utilized: seismic refraction and seismic reflection.

Seismic refraction. The field measurements are made with seismic spreads consisting of twelve or twenty-four geophones equally spaced at distance d along a straight line. Usually five explosions are made for one spread, all in line with the geophones. One of the shotpoints is located at the centre between geophones 6 and 7 (or 12 and 13), the others being outside the spread at a distance $d/2$ and $11d$ (or $23d$) from geophones 1 and 12 (or 24). Figure 9.1.2.2a is an example of a recording.

Let us consider three horizontal layers with velocities V_1, V_2, V_3 such that $V_1 < V_2 < V_3$, a seismic spread, and a shotpoint E placed at a distance d from the first geophone (Fig. 9.1.2.2b). The first recorded wave to reach the geophones close to the shotpoint is the direct wave propagating along the soil surface. The more distant geophones are first reached by waves refracted along deeper and deeper interfaces. The farther the geophones are from the shotpoint, the deeper the interfaces where the waves are refracted. In seismic refraction the depth of investigation H is controlled by the length L of the layout of the geophones; the value of H being about $L/5$.

A time-distance curve is built from each record, plotting the arrival time of refracted waves against the distance between the shotpoint and the geophones. This curve has a number of segments equal to the number of layers and the slopes of the segments give the reciprocals of the velocities for the various beds. From the intercepts T_1, T_2 of segments II and III, it is possible to calculate the thicknesses m_1 and m_2 of the two top layers using the two following formulae:

$$m_1 \cdot \frac{T_1}{2} = \frac{V_1}{\cos i_{1,2}} \quad \text{or} \quad m_2 = \frac{T_1}{2} V_2 \tan i_{1,2}$$

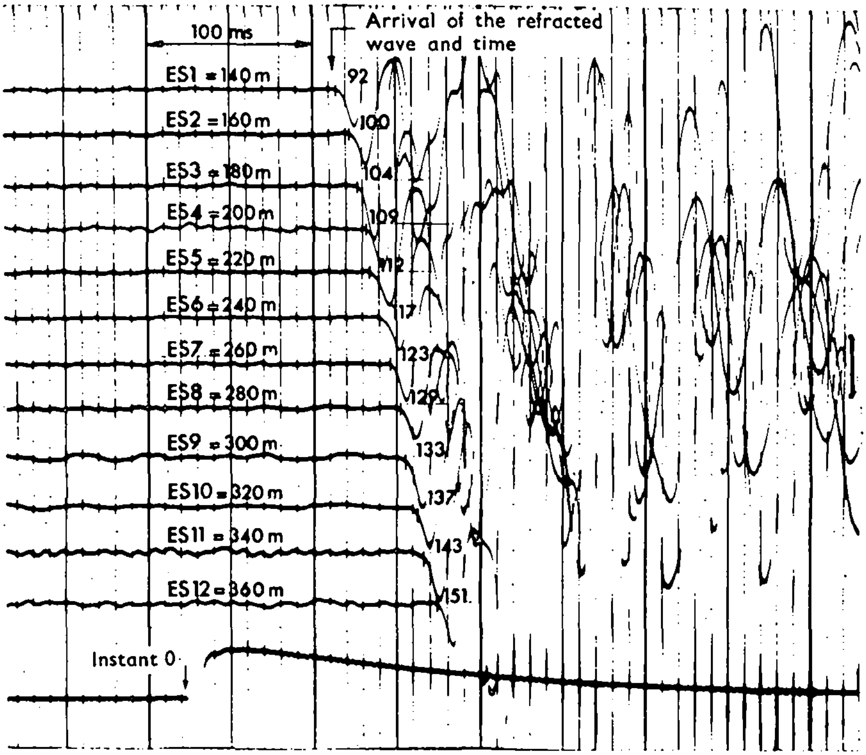


FIG. 9.1.2.2a. Example of a seismic refraction recording. (Lerici.)

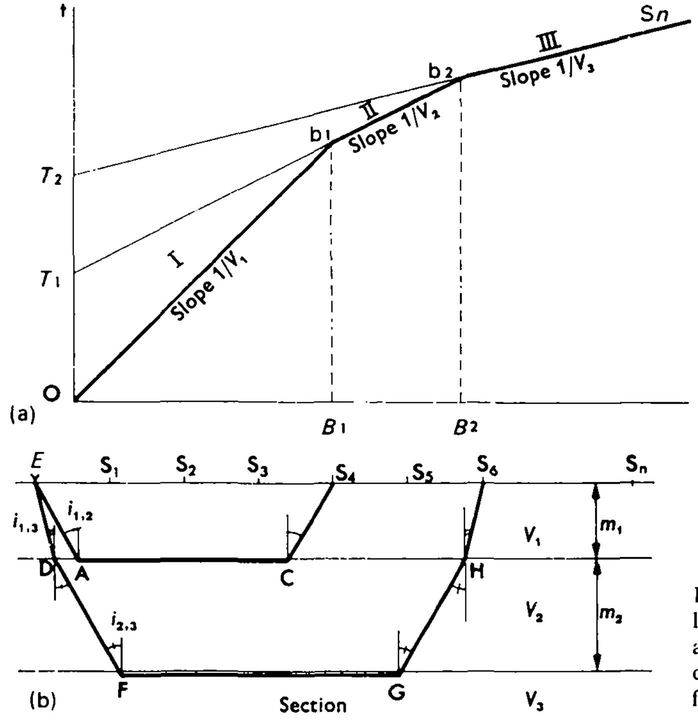


FIG. 9.1.2.2b. Three layers separated by parallel planes: (a) time-distance curve; (b) refracted-ray paths.

because $V_1 = V_2 \sin i_{1,2}$,

$$m_2 = \frac{T_2 - T_1}{2} \cdot \frac{V_2}{\cos i_{2,3}} \text{ or } m_2 = \frac{T_2 - T_1}{2} V_3 \tan i_{2,3}$$

because $V_2 = V_3 \sin i_{2,3}$.

In the usual case of inclined interfaces, the slope of the segments of the time-distance curve permits calculating an apparent velocity of the layers but not the true velocity. Apparent velocities obtained from direct shotpoints (situated alongside the first geophone) and reserve shotpoints (situated alongside the last geophone) have to be combined to calculate the true velocities. The two apparent velocities to be combined must refer to the same portion of the basement, and only distant external shotpoints can provide them. That is why direct and reverse distant shotpoints are necessary.

Thin layers and layers with a velocity less than that of the overlying bed cannot be observed on the time-distance curves. Thin layers cause negative errors and low velocity layers cause positive errors both in depth and thickness calculations.

Seismic refraction surveys consist either of soundings or profiles. A seismic sounding corresponds to an isolated spread with five shotpoints. A profile is made up of several spreads in line, the last geophone of spread n being the first one of spread $n + 1$.

The arc shooting technique utilizes refracted waves, the shotpoint being at the centre of a circular arc formed by the geophones. Arc shooting is seldom employed because it is inaccurate when there are more than two layers.

Seismic reflection. In seismic reflection prospecting, any interface between geological beds with different velocities acts as a mirror in reflecting elastic waves back towards the surface, whether the higher velocity bed is above or below. The higher the velocity contrast between the two beds, the higher the amount of reflected energy.

Consider a geological section of four horizontal layers and a seismic line consisting of a shotpoint and some regularly spaced geophones (Fig. 9.1.2.2c). The travel time-distance equation is a hyperbole of the form:

$$\frac{t^2}{\frac{4H^2}{V_m^2}} - \frac{x^2}{4H^2} = 1$$

where:

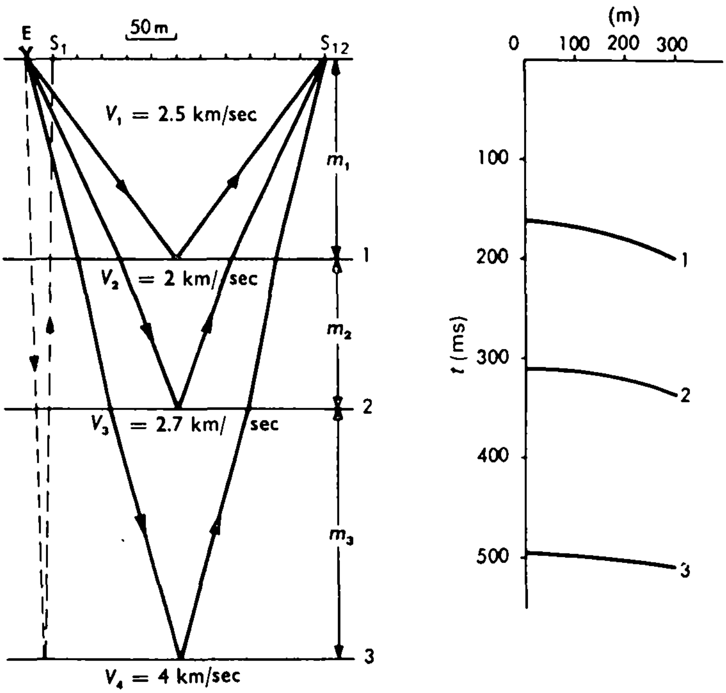
H = depth to reflector;
V_m = average velocity of the layers overlying the reflector;
x = distance between a shotpoint and a geophone.

If the one-way time is t_s, the formula for the receiver nearest to the shotpoint, for which x^2 can be neglected, is: $H \simeq t_s V_m$.

Figure 9.1.2.2d is a recording of a seismic reflection obtained with a shotpoint placed in the centre of the geophone spread.

In seismic reflection prospecting the depth of investigation does not depend upon the length of the seismic spread. It depends upon the amount of energy reflected by deep interfaces and upon the sensitivity of the equipment.

Seismic reflection is the method which in the last two decades has made the most progress, taking advantage of magnetic recording, digital systems and optical processing. It nearly always gives the best and most accurate geological sections. Often, ten or more contacts can be followed with reasonable precision. With other methods rarely more than four can be followed, and the precision decreases rapidly with the rank of the marked bed. However, if seismic reflection is by far the most common method employed in petroleum exploration, it is almost never used in hydrogeology. Seismic refraction is preferred for the three following reasons:



Reflected-ray paths

Time-distance curve

FIG. 9.1.2.2c. Seismic reflection: reflected-ray paths and time-distance curve.

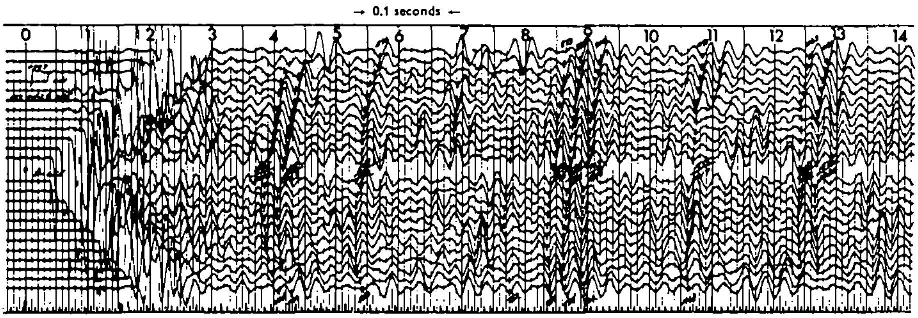


FIG. 9.1.2.2d. Seismic reflection galvanometric recording with indication of the arrivals of reflected waves.

1. Seismic reflection is poorly adapted for shallow studies and gives good results only for depths greater than 200 m. The reflections from shallow contacts arrive at the geophones before the direct waves do so and before the waves refracted along the base of the weathered zone have died down, which is the reason they cannot be detected. Periodically, systems are proposed to adapt the method for shallow exploration (Tsi Meidav, 1969; Leandri and Collin, 1972).
2. Seismic reflection is much more efficient when used for long continuous profiles rather than for small scattered profiles comparable to the refraction soundings.

3. In hydrogeology the price of even the smallest reflection crew is much more than that of a refraction crew, mainly because it is necessary to bury the explosive charge under the dry and loose superficial sediments, which means using a drilling rig instead of a hand auger.

Wadi Azum Valley (Sudan). During the joint project of the Sudan Government and the Food and Agriculture Organization of the United Nations for the development of the Jebel Marra area, a geophysical survey was made in the Wadi Azum Valley to determine the volume of the alluvial reservoir.

In some sections of the valley, the Holocene sands and gravels have the same resistivity as the bedrock of schist, gneiss and granite. Seismic soundings were made because the velocities of the elastic waves in the dry alluvium, saturated alluvium and bedrock are very different: respectively 0.3 km/sec., 1.5 to 2 km/sec. and 3.7 to 5.2 km/sec.

Figure 9.1.2.2e reproduces the time-distance curves corresponding to seismic sounding No. 1. Figure 9.1.2.2f gives a cross-section of the valley as worked out from seismic soundings.

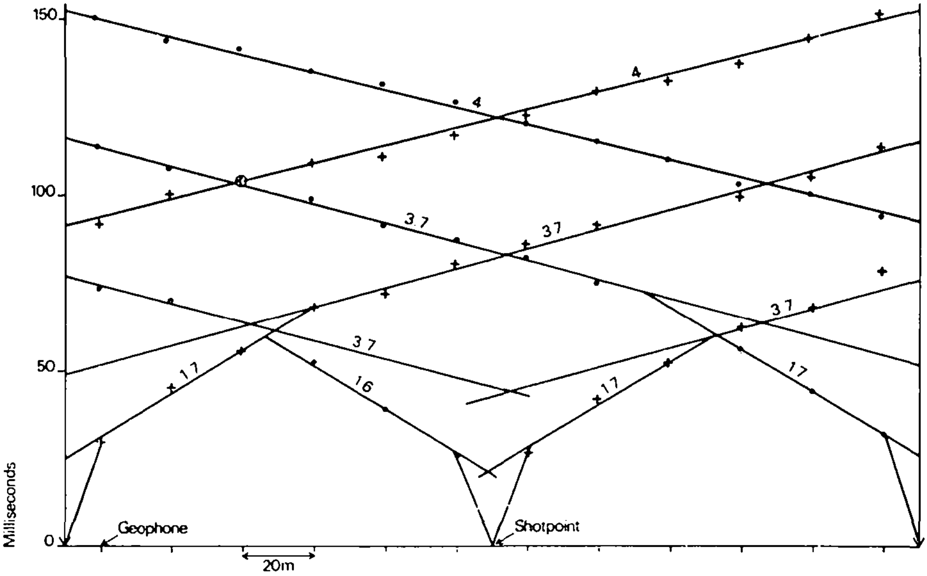


FIG. 9.1.2.2e. Wadi Azum Valley: seismic refraction sounding No. 1, time-distance curves (velocity in km/sec.). (FAO.)

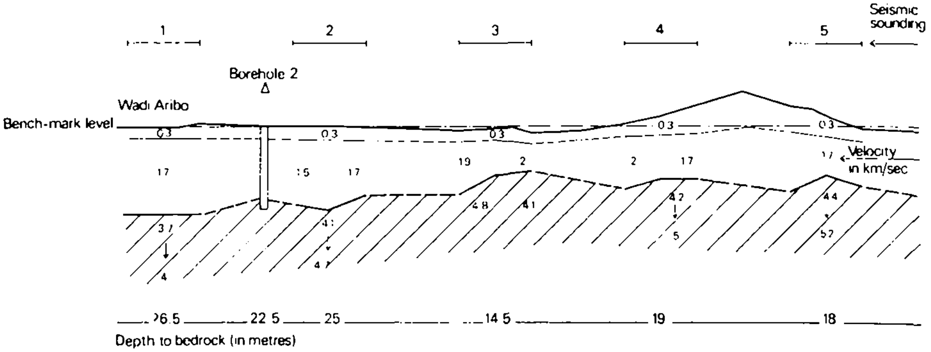


FIG. 9.1.2.2f. Wadi Azum Valley: results of the seismic refraction soundings along section Z13. (FAO.)

Kuwait. A seismic reflection survey was made for the Department of Water and Gas of Kuwait to discover fractured areas in a limestone formation 200 to 500 m deep.

Figure 9.1.2.2g reproduces a part of a corrected variable density record section. On the left side of the recording, the seismic markers are interrupted because the reflected waves are replaced by diffracted ones. The zone where this occurs corresponds to a fault without major displacement.

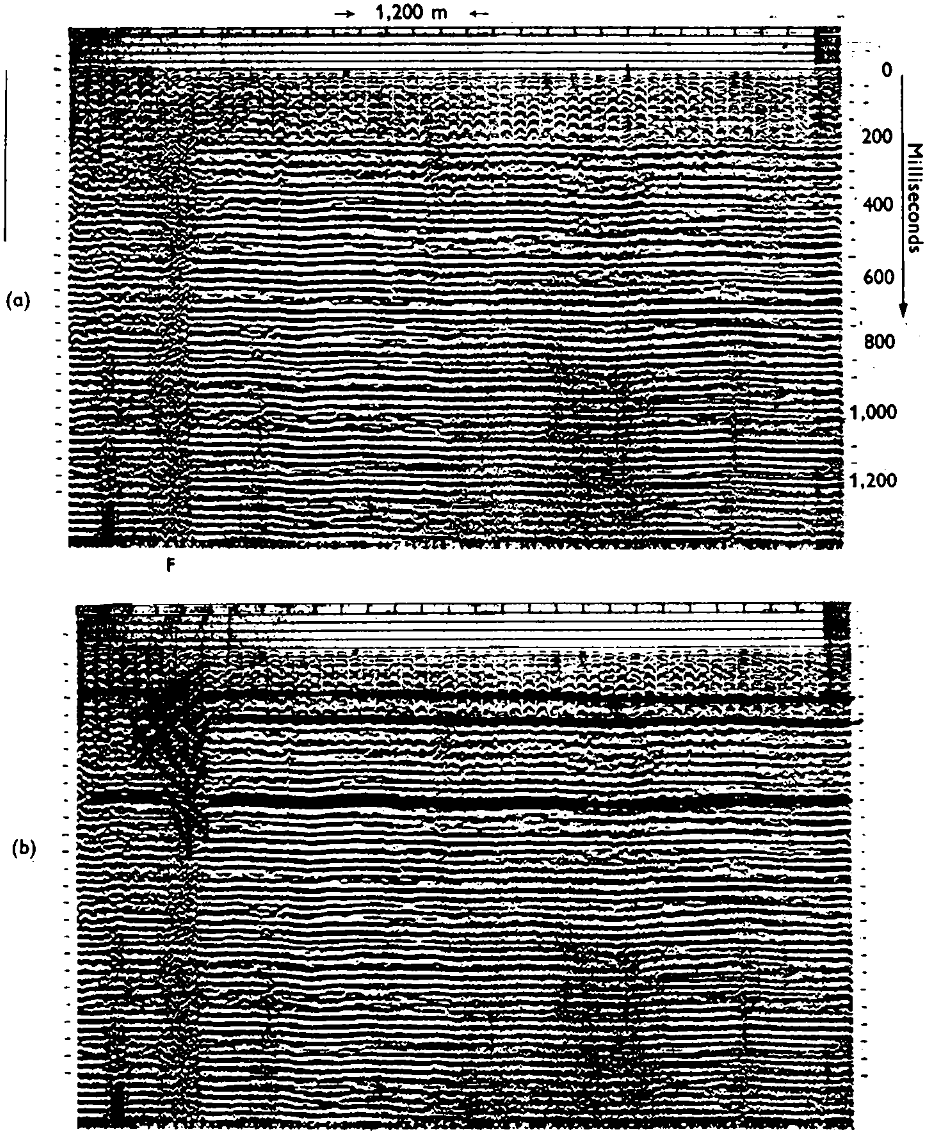


FIG. 9.1.2.2g. Seismic reflection, variable density record section: (a) section without interpretation; (b) section with interpretation. The strip on the left side where the markers are interrupted corresponds to a fault. (Department of Water and Gas, Kuwait.)

9.1.2.3 Gravity prospecting

The weight of a body is equal to the gravitational attraction of the earth. This is Newton's Law, and the formula is:

$$mg = \gamma \frac{Mm}{r^2}$$

where: m and M are the masses of the body and the earth respectively; r is the radius of the earth; γ the universal gravitational constant; and g is the acceleration due to gravity. The acceleration g is expressed in cm/sec^2 or gals, but the practical units used in prospecting are the milligal and the gravity unit (g.u.) which is equal to 0.1 milligal.

If the earth were homogeneous, perfectly spherical and motionless in empty space, g would be the same over the whole surface. In actual fact, g varies appreciably from point to point on the earth. In gravity prospecting the g variations are measured, and corrections are applied for the attraction of celestial bodies and the rotation and shape of the earth, in order to determine the anomalies caused by ground density variations. Table 9.1.2.3 gives the density of the more common rocks.

TABLE 9.1.2.3. Density of the more common rocks

Rock	Density	Rock	Density
Dry loams	1.4–1.8	Sandstones	2–2.6
Saturated loams	1.8–2.2	Limestones and dolomites	2.2–2.8
Dry sands and gravels	1.4–2.2	Schists	2.4–2.8
Saturated sands and gravels	1.8–2.4	Granites	2.6–2.8
Clays	1.7–2.5	Gneiss	2.6–3
Massive gypsum or salt	2.1–2.3	Lavas	2.8–3
Marls	1.8–2.6	Anhydrite	2.8–3

Variations in gravitational attraction are measured with gravimeters, the operating principle of which is very simple. The weight of a small mass is compensated by a spring, and the length variations of the spring are observed.

Bouguer anomaly maps, residual anomaly maps and vertical gradient maps are established. The Bouguer anomaly map gives the difference of the corrected values of g between each station and a reference station. The residual anomaly is the difference between the Bouguer anomaly and the large anomaly which is associated with the crystalline rock basement. The vertical gradient map is a map of the first derivatives dg/dz, which emphasizes local anomalies.

In hydrogeology, gravity prospecting is mainly used for: (a) determining the shape of a bedrock surface beneath unconsolidated sediments covering large flat plains; (b) locating karstic or fractured limestones and underground rivers. High precision surveys permit considering anomalies of an amplitude of 0.02 milligal as being significant (Neumann, 1972). A fault which, in a strip 10 m wide, causes an increase of 10 per cent in the porosity of limestones or lavas, gives an anomaly of at least 0.05 milligal if the fractured rocks are less than 8 m deep. A cylindrical channel, 6 m in diameter, will correspond to an anomaly of 0.05 milligal if its axis is 17 m deep.

9.1.2.4 Thermometry

Geothermic prospecting is used in hydrogeology in the following cases: (a) location of the original spring or the rock reservoir of thermo-mineral waters; (b) direct detection of underground water less than 20 m deep in alluvial deposits or high porosity rocks.

The direct detection of underground water is possible because the specific heat of water is equal to 1 whereas that of the majority of dry geological formations is 0.2–0.3. For this reason, high porosity aquifers located at shallow depth will heat and cool with the seasons much less rapidly than impermeable formations.

In remote areas where drilling is very expensive, the association of geothermic prospecting with a resistivity or a seismic refraction survey may avoid the siting of boreholes in dry reservoirs.

The example of Figure 9.1.2.4 published in 1968 by Cartwright, is from Illinois (United States). It shows that the thermal variation related to the ground water reaches −1.5° C in summer, +1° C in winter. The water table is 4.5 to 6 m deep and the thermal measurements have been made in small holes, 0.6 to 1 m deep.

Kremar and Masin (1970) in Czechoslovakia have successfully used geothermic surveys to find shallow water-bearing cavities and faults.

9.1.3 *Airborne geophysical prospecting*

9.1.3.1 **Infra-red thermometry**

Among the so-called remote-sensing methods developed in the last ten years, infra-red thermometry is the most used in hydrogeology.

The intensity of the infra-red radiation emitted by a body depends upon its emissivity coefficient and the fourth power of its absolute temperature. As a first approximation, it is assumed that the emissivities of rocks and water have the same order of magnitude and that the intensity of the infra-red radiation from the earth's surface depends essentially on the temperature of the different substances. Infra-red thermometry consists of the measurement of the temperature of the earth's surface through the detection of the infra-red radiation.

The detectors are special crystals which translate radiation differences into weak electrical signals which are amplified and then used to modulate a variable-intensity light source. The final document is an infra-red image which is similar to an aerial photograph. It is important to remember, however, that conventional photographs represent variations in reflectivity whereas infra-red images give the pattern of temperature distribution. In fact, most infra-red thermic flights take place before dawn to eliminate the effects of any reflected energy in the infra-red spectrum. On the infra-red images, the colder areas normally appear darker than the warmer ones, although the opposite may be the case, depending on the processing technique.

To determine the temperature corresponding to each shade of grey on the infra-red image, internal calibration sources or auxiliary radiometers are used. Sophisticated processing allows the drawing of 0.1° C isotherms.

The infra-red spectrum lies between the wavelengths of 0.76 micron and 1,000 micron. Infra-red imagery is collected in either the 3 to 5 micron or the 8 to 14 micron regions where the atmospheric energy attenuation is the lowest.

During the flight a rotative mirror scans adjacent strips of ground perpendicular to the aircraft's flight path (Fig. 9.1.3.1a). The maximum view angle is 80–100°.

Infra-red thermometry can be employed in hydrogeology: (a) to discover small and shallow water basins in arid regions; and (b) to locate points of discharge of ground water flowing out of karst channels or fossil valleys into the sea, lakes or rivers.

The infra-red image reproduced in Figure 9.1.3.1b was obtained during a survey carried out for a joint project of the Lebanese Government and the Food and Agriculture

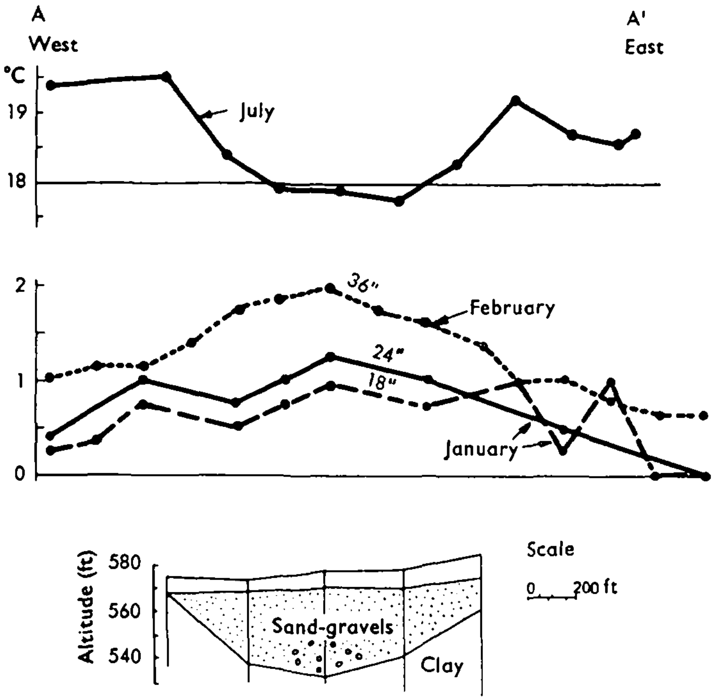


FIG. 9.1.2.4. Finding an aquiferous formation by temperature measurement. (USGS, Illinois.)

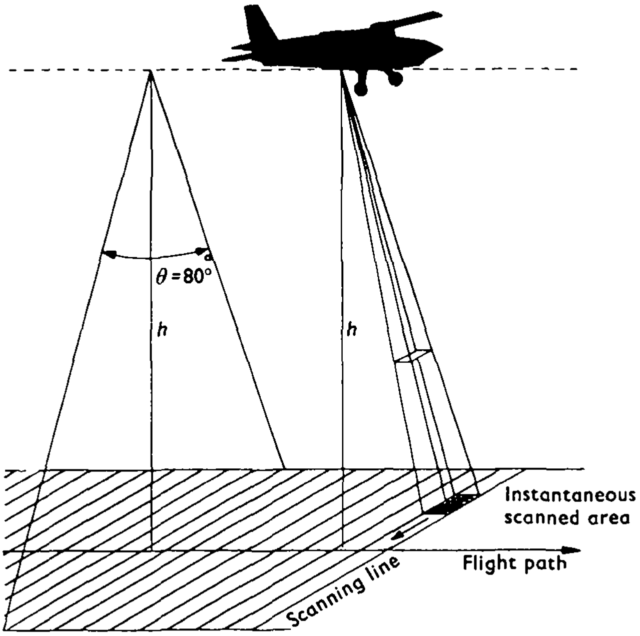


FIG. 9.1.3.1a. Infra-red scanning.

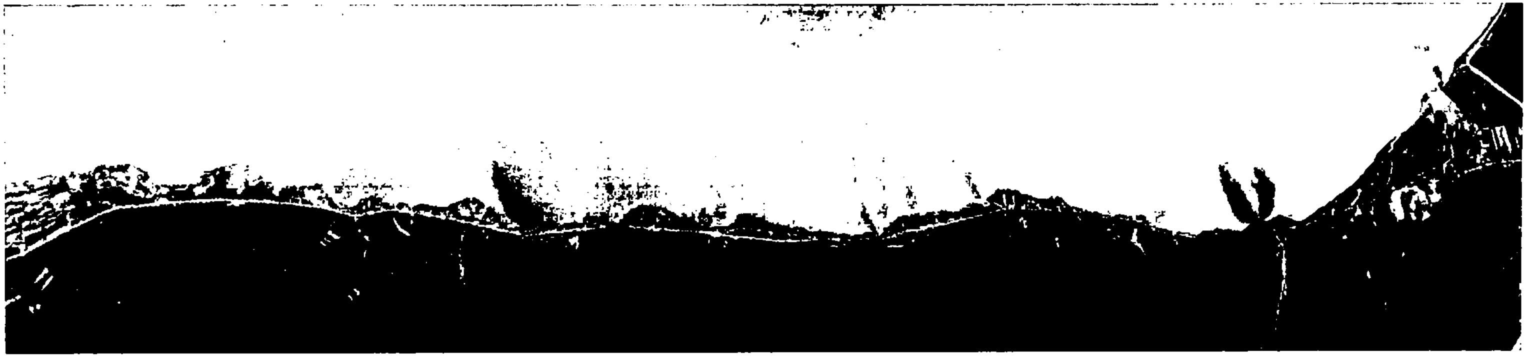


FIG. 9.1.3.1b. Example of black and white thermal imagery. (FAO.)

Organization of the United Nations. On this image the horn-shaped anomaly is particularly interesting because it does not correspond to a surface-water runoff, and it may represent the underflow of the near-by river.

9.1.3.2 **Radio-wave methods**

Different methods employing radio waves for the measurement of ground resistivity have been studied. The depth of investigation is very limited, especially if a conductive soil exists. Even using the VLF band (20 KHz) under favourable conditions, it does not exceed 60 m.

Recently, a new airborne operative method called E-Phase R system has been developed. Apparent ground resistivity is calculated from the complex ratio of the horizontal electric field component to the vertical electric field component of the radio waves. Military and amateur transmitters broadcasting in the 15 to 1,500 KHz band and located far from the area to be surveyed are used. If two or three radio stations with different frequencies are chosen, measurements at two or three different depths can be made simultaneously.

The radio-wave method can be used in remote and arid countries to find small and shallow water basins. For example, subsurface reservoirs which develop in weathered areas of crystalline bedrock can be detected as conductive anomalies.

9.2 Geophysical logging of boreholes

9.2.1 *Introduction*

Numerous techniques have been developed for the geophysical exploration of probing to increase the precision of interpretations based on the geologist's examination of cuttings returned by the drilling fluid, and to avoid the high cost of core samples.

These techniques were first developed for the needs of petroleum exploration but were later extended to ground-water exploration. In hydrogeology only the simplest and least expensive methods are used, except in special cases.

The interpretation of the geophysical logs differs somewhat in hydrogeology and in petroleum exploration. The hydrogeologist is concerned mainly with fresh water or fresh- and salt-water interfaces instead of oil-gas-brine interfaces; moreover, boreholes drilled for water are not usually deep enough to reach high-temperature strata.

The geophysical probing of boreholes or geophysical logging is the measurement and interpretation of the following properties of a sequence of rock strata and their contained fluids: resistivity, natural or induced radioactivity, seismic velocity, temperature and dip. Other measurements made in boreholes concern the hole itself or phenomena associated with it and include diameter, mud resistivity, spontaneous potential, water flow.

Well logs serve to identify the rock types; they establish the position in depth of boundaries between rock strata, and they assist in geologic correlations between boreholes. They also allow a preliminary evaluation of water salinity and of the total porosity and permeability of aquifers. Finally, they help in the interpretation of geophysical surface surveys.

As a general rule, the measurements are recorded continuously, the film or paper speed of the recorder being synchronized with that of the sonde in the hole. Except in particular cases, the recording is done with the sonde coming up the hole in order to keep the cable taut. The speed of the probe varies and depends upon the equipment and the parameter being measured. It is usually between 6 and 30 m per minute. All measurements may be made in uncased boreholes, but in cased boreholes only radioactive measurements can be made (temperature logs and water-flow measurements can also be recorded if the casing is perforated).

Interpretation of the well logs can sometimes be done immediately, but often it requires the use of master curves. When many well logs are available, they may be mechanically processed to give a complete quantitative interpretation.

9.2.2 *Electrical logging*

9.2.2.1 **Spontaneous potential log**

The spontaneous potential log (SP) corresponds to the recording of the potentials occurring naturally in a borehole. These potentials are mainly caused by the differences in salinity between the drilling mud, the water in the aquifers and the water contained in the clays. Different SP curves can be obtained from the same borehole by changing the mud salinity.

The clayey formations in a borehole usually define a base line from which the potential of permeable formations can be determined (Fig. 9.2.2.1). When both the mud and the ground water can be considered as NaCl solutions, the spontaneous potential of an aquifer equals

$$SP_{\mathrm{millivolts}} = K.\ \log \frac{\rho_m}{\rho_w}$$

where: ρ_m = mud resistivity; ρ_w = ground-water resisitivity and **K** = temperature coefficient.

In most cases, if an aquifer gives a positive SP, the water resistivity is higher than that of the mud.

9.2.2.2 **Resistivity logs**

The resistivity measured in a borehole is an apparent resistivity which may deviate considerably from the true rock resistivity. It is influenced by the configuration of the measuring electrodes, the resistivity of the drilling fluid, the diameter of the borehole, the thickness of the rock strata and the depth of infiltration of the drilling fluid into the walls of the borehole.

In hydrogeology, the most commonly used measuring devices are the single electrode, the normal and lateral devices (Fig. 9.2.2.2).

The single electrode has a very limited depth range. It accurately indicates the boundaries of a rock stratum but the apparent measured resistivities differ greatly from the true resistivities.

Generally two normal devices are used simultaneously, a short one with an electrode separation AM = 16 inches and a long one for which AM = 64 inches. The radius of investigation is about 2 AM which makes it about 0.8 m for the short normal and 3.2 m for the long normal. The short normal device accurately indicates the boundaries of the rock stratum and, for thick permeable formations, it gives an apparent resistivity which is close to the true resistivity of the zone invaded by the drilling mud. The long normal device does not react to the thin resistive beds (thickness less than 1.6 m) and does not indicate boundaries with much accuracy, but it does give the true resistivity of the thick beds (at least 1.6 m thick for conductive beds, 8 m for resistive beds). The combination of short and long normal devices provides all the necessary information.

For the lateral device commonly used, the interval AO and the radius of investigation equal 5.4 m. The advantage of the lateral device over the long normal device is in locating thin resistive beds, but the lateral log is more difficult to interpret.

Examples of short normal, long normal and lateral logs are reproduced in Figure 9.2.2.1.

Microdevices and focusing devices have been developed but are rarely used in hydrogeology. The microdevices have a shallow penetration and are mainly used to measure the resistivity of the invaded zone and to delineate thin permeable beds. Focusing devices increase the investigation depth and minimize the effects of the borehole and of the adjacent formations.

9.2.2.3 **Mud resistivity measurement**

For a quantitative interpretation of the resistivity logs, it is necessary to know the mud resistivity of the drilling mud which can be determined from a mud sample.

Recordings of the mud resistivity according to depth are sometimes made for the purpose of finding zones where water enters or escapes from the hole.

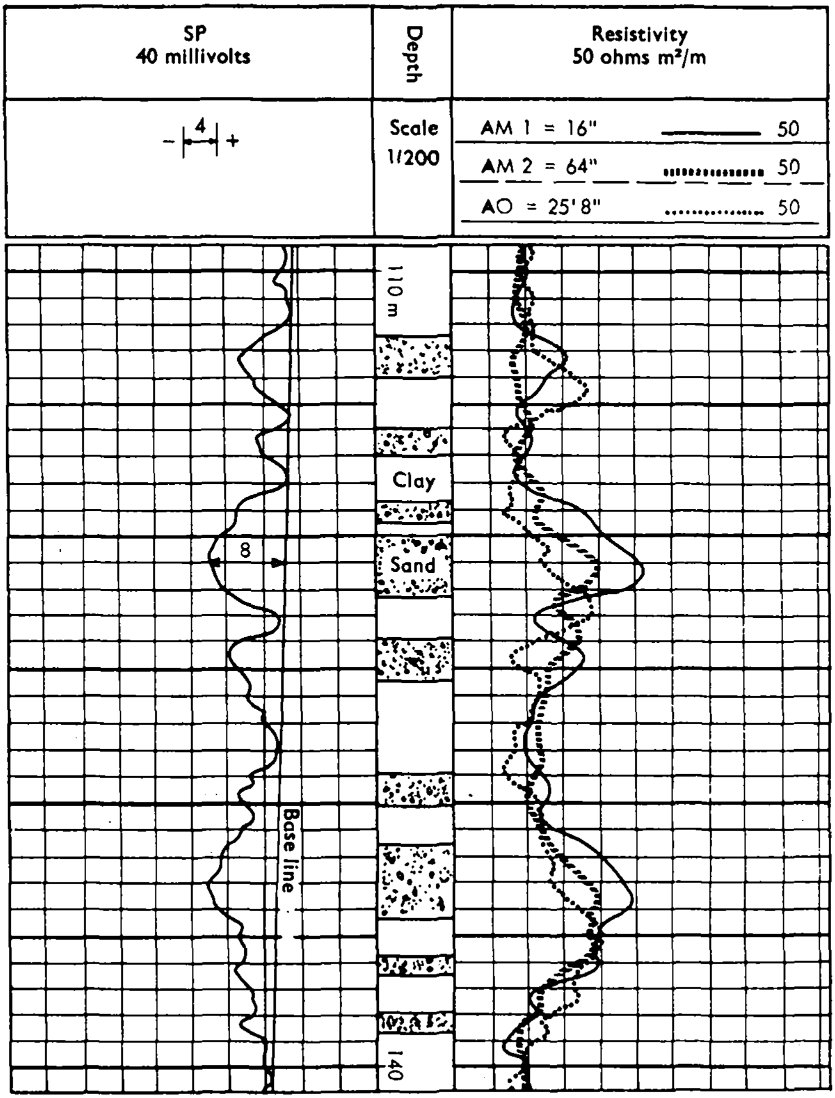


FIG. 9.2.2.1. Spontaneous potential and resistivity logs (AM 1 = short normal device; AM 2 = long normal device; AO = lateral device). (SPE.)

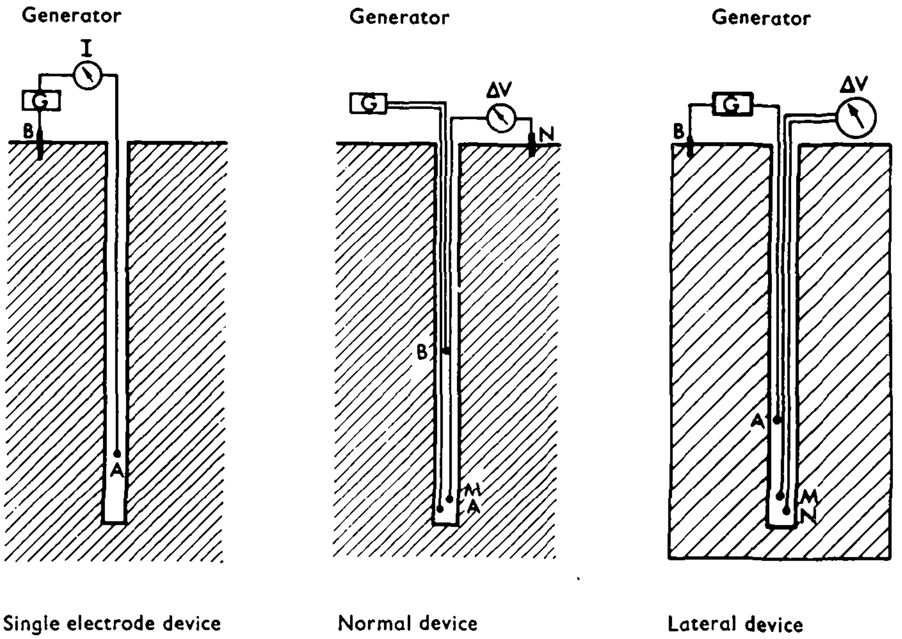


FIG. 9.2.2.2. Devices used for resistivity logging.

9.2.2.4 **Fluidconductivity log**

The ionic content of ground water in a well can be assessed by measuring fluid resistivity (at a short electrode spacing) or by determining directly electrical conductivity by special electrode assemblies. Electrical conductivity (EC) increases with increasing ionic concentration, so that these types of logs are of particular importance where adverse mineralization or saline intrusion must be monitored (Ineson and Gray, 1963). These types of logs can be used for identifying aquifers and determining ground-water seepage rates (Ogilvi and Fedorovich, 1966).

9.2.3 *Caliper log*

The caliper log measures the bore diameter by means of a sonde having three spring-loaded arms which keep them in contact with the sides of the hole. The expansion or compression of the arms causes a piston to move along the axis of the sonde which, in turn, varies the induction between two coils or the resistance of a potentiometer. The resulting electrical current is proportional to the diameter of the hole.

For the precise interpretation of most well logs, it is useful to know the diameter of the borehole. The caliper log also serves: (a) to identify certain strata which generally coincide with increases in the diameter, for example clay formations, fractured zones (see Figure 9.2.3) and sand beds in consolidated rocks; (b) to calculate the volume of cement to inject behind the casing; (c) to evaluate results of explosive shots used for shattering hard rock strata or boulders—for this two logs are necessary, one before and one after the shot.

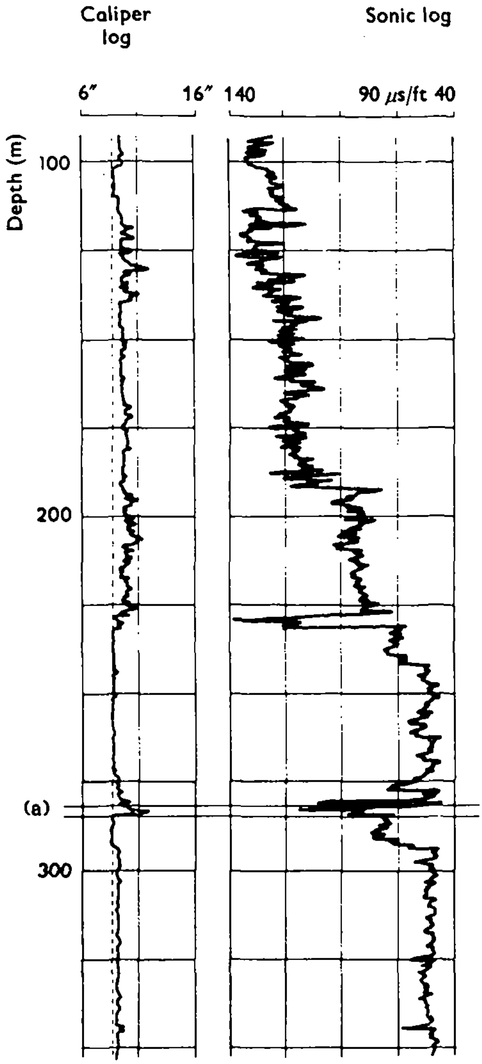


FIG. 9.2.3. Caliper log and sonic log showing a fractured zone (a) in dolomites.

9.2.4 *Sonic logging (velocity log)*

Sonic logging measures the speed of sound waves passing through the layers penetrated by a borehole. The logging records the minimum time it takes ultrasonic waves to travel the distance of several tens of centimetres which separates two receivers. This time is the reciprocal of the required velocity.

Sonic logs are particularly useful for stratigraphic correlation in different bores for the detection of fractured zones and for the calculation of the total porosity of consolidated formations.

Figure 9.2.3 shows how a fractured zone in dolomites affects the sonic log.

9.2.5 *Temperature log*

The main use of temperature logs is in finding permeable strata. Figure 9.2.5 shows temperature logs taken at different times after stopping mud circulation in a borehole penetrating two sandy formations. The temperature of the mud is warmer than that in the shallow formation and colder than the deeper one. To obtain reliable results, the log should be recorded a few hours after stopping the mud circulation.

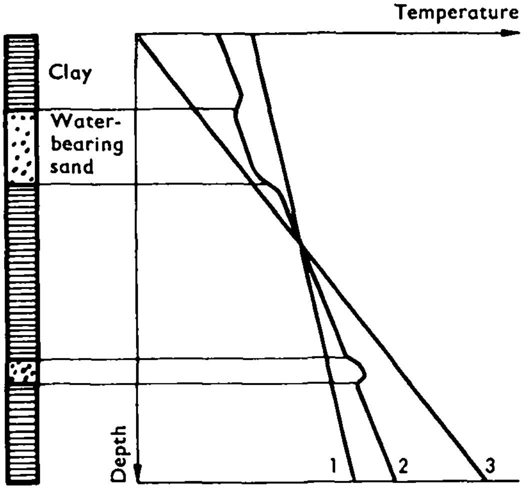


FIG. 9.2.5. Temperature log showing water-bearing sands: curve 1, immediately after stopping mud circulation; curve 2, a few hours after stopping mud circulation; curve 3, a few days after stopping mud circulation.

9.2.6 *Water flow log or flowmeter (F)*

Varying velocities of ground-water movement can be studied in a well, due to both incremental inflow and decremental outflow. The movement can be established by following the dispersal of injected tracers or by instruments placed in the well to measure directly or indirectly ground-water velocities.

Ground-water movement is measured by specially developed flowmeters, similar to miniature current meters and capable of measuring very low velocities, or by thermometric instruments. The movement will be either upwards or downwards within the well. The temperature profile can be studied during abstraction within the well or in an adjacent well, the resulting profile not uncommonly being significantly different from the natural thermal profile.

Flowmeter diagrams make it possible to determine the hydrodynamic characteristics of aquifers (Grinbaum, 1965; Gershanovich, 1966).

9.2.7 **Dip meter**

With the help of a single probe, in which a borehole deviation-measuring device is combined with three resistivity-measuring electrodes in contact with the wall of the borehole, it is possible to measure the strike and dip of individual beds.

9.2.8 *Nuclear logging*

Nuclear well logging is based on the measurement of natural or artificially induced radioactivity during the introduction of a logging probe in the well. All rocks contain small amounts of natural radioactive elements, like uranium and thorium, with the products of their decay, as well as the radioactive isotope of potassium ^{40}K. When used in a bore, the logging probe is screened against cosmic radiation and the background of the gammaradiation counters in the probe is so low that even a very weak radiation from rocks can be detected and recorded with a rather high accuracy. Such measurements are known as natural gamma logging.

Geochemistry permits the determination of the average content of radioactive elements in different types of rocks (Table 9.2.8).

TABLE 9.2.8. Content of radioactive elements in some types of sedimentary rocks (Filippov, 1962)

Type of rock	Ra (pg/g)	U (µg/g)	Th (µg/g)	Th/U	^{40}K (g/g)
Limestone	0.5	1.50	0.5	0.33	0.001
Dolomite	0.11	0.3	—	—	—
Quartzite	0.54	1.6	—	—	--
Sandstone	0/1.5	up to 4.0	5.0	1.2	0.004
Clay shale	1.09	3.0	10.0	3.3	0.008
Clay	1.3	4.3	13.0	3.0	0.011

Using the natural gamma, it is generally possible to distinguish between certain rocks in the stratigraphy profile and to determine the depth of their occurrence.

Modifications of the nuclear log are the gamma-gamma (or density) and the neutron logs. Their concept is based on how gamma radiation and the neutrons react with the atoms of rocks. In this case, an artificially prepared source of gamma quanta or neutrons is attached to the housing of the gamma probe, properly screened to prevent their being recorded directly by the detector in the probe.

Under these conditions, gamma quanta and neutrons enter the formation where they either undergo scattering or, in the case of neutrons, induce new radiation. A radiation detector placed in the axis of the borehole will record them and send the information to the surface.

These two variations of nuclear logs can yield additional information concerning the density of rocks and the moisture content. For more detailed information on nuclear interactions and further discussion of the physical principles of nuclear logs, the reader is referred to International Atomic Energy Agency (IAEA), 1968, 1971.

Equipment. A nuclear well-logging system is shown diagrammatically in Figure 9.2.8a. When lowered into the borehole on a logging cable, the probe contains radiation detectors, usually of the Geiger-Müller type, scintillation or neutron proportional counters, and also an electronic section (amplifier, high voltage source, pulse formation circuit, etc.). They are enclosed in a pressure-proof housing into which nuclear radiation can easily penetrate.

For shallow wells, portable equipment with a manually operated winch can be used. If the depth is more than 50 m, it is more convenient to use a special truck, equipped

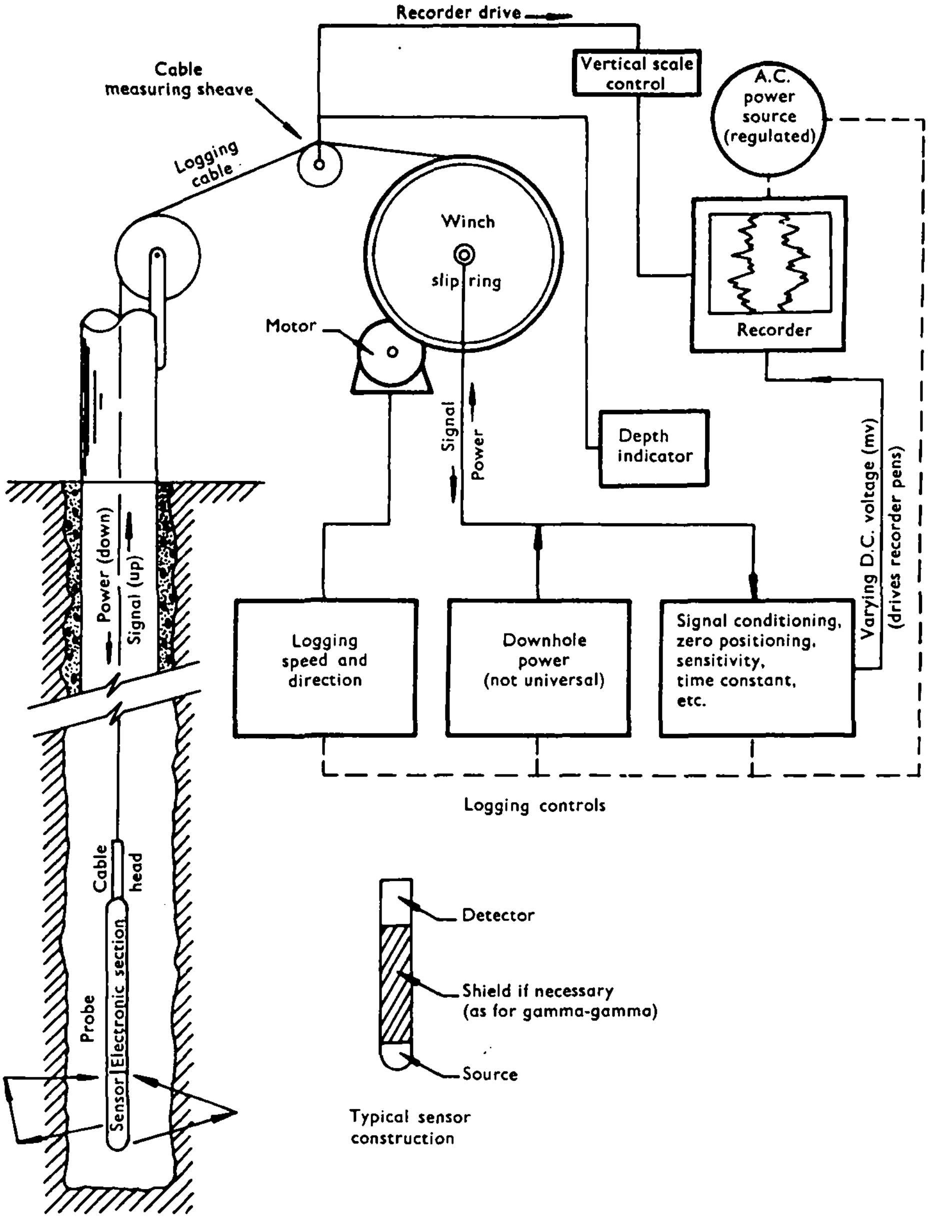


FIG. 9.2.8a. Basic components of geophysical well-logging equipment. (Keys and MacCary, 1971.)

with a permanently installed cable winch, control units, rate-meter and logging recorders, as well as equipment for other non-nuclear logs, and power supply.

The most efficient conditions for using the nuclear log are when both the borehole and the logging probe have a small diameter. If this is not the case, a special spring must be used to press the probe on to one wall of the well.

Another method for introducing the logging probe into the ground is the following. Iron tubes are driven by a hydraulic device (Ferronsky, 1969) or a vibration hammer (Borowczyk, 1970) into the unconsolidated material without prior drilling of a borehole (Fig. 9.2.8b). During the process of emplacement, however, even a hole which penetrates only a few tens of metres may yield some additional information about the physical parameters of the soil.

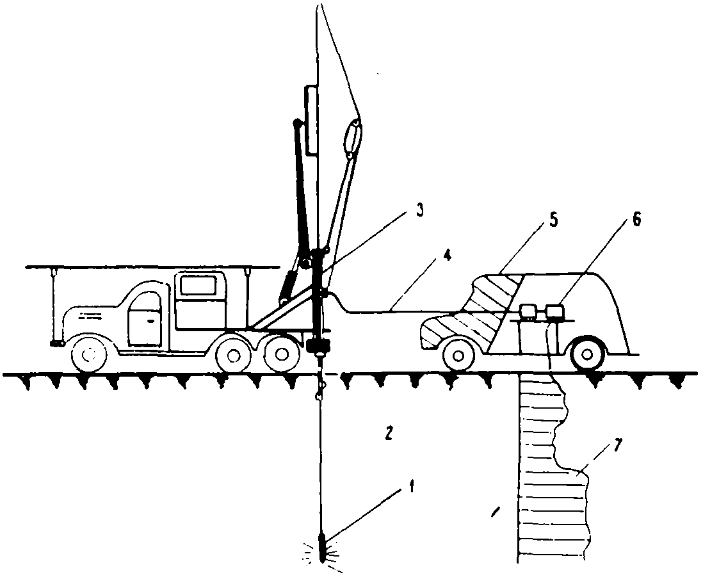


FIG. 9.2.8b. Diagram of installation for the investigation of saturated and unsaturated zones, in combination with well sinking by the method of embedding. 1. Logging probe; 2. Shaft; 3. Apparatus for embedding the shaft; 4. Cable; 5. Equipment laboratory; 6. Recording apparatus; 7. Log.

Highly consolidated strata, cobbles or pebbles and cemented zones may prevent penetration.

The vibration penetration method permits the installation of a small-diameter casing with a screen so that the hole can be used for certain hydraulic tests and for observation purposes.

Safety. Excessive nuclear radiation can endanger health. In logging instruments, radioactive sources are in sealed capsules and only irradiation of the operators need be considered (radioactive contamination is not very probable under such conditions). The permissible gamma radiation dose is 170 millirems/year to the whole body. The permissible neutron dose is 17 fast neutrons/cm^2. sec. or 670 thermal neutrons/cm^2. sec. to the whole body. People working with radioactive sources should be equipped with suitable film or other radiation detectors in order to maintain an exposure record. Special authorization is required in most countries for handling radioactive sources.

9.2.8.1 **Natural gamma-ray log**

The natural gamma log is most commonly used to determine the clay content in a lithological profile, as clays and clay shales contain more U, Th and Ra than other sedimentary rocks (Table 9.2.8). This is the most common case for their application. Crystalline igneous rocks, and especially acidic rocks, have an average content of radioactive elements comparable to that for clays, but higher than sedimentary rocks.

After proper calibration of the logging probe, a quantitative interpretation of the gamma log may be made. The units used in such a case are: counts per second, microroentgens per hour, or so-called API units (American Petroleum Institute units). This method appears to have a valuable future. Keys and MacCary (1971) have written: 'The most important and, to date, little used function of water-well logs is to quantify geohydrologic parameters.'

9.2.8.2 **Gamma-gamma logging**

In principle this involves the recording of rock-scattered radiation from a gamma source which is carried by a logging probe as it moves down the well. The scattered gamma radiation is related functionally to the density of the surrounding rocks, so that the method records the rock density as a function of depth in the well. The log is normally used in the borehole profile to identify rocks of higher than average density (for example anhydrite, barite), or lower than average (for example lignites, coal, clays).

After calibration, and preserving conditions of similarity between the borehole and the calibration model, quantitative interpretation can continue to be made once the probe has been calibrated, so long as conditions between the borehole and the calibration model remain comparable.

9.2.8.3 **Neutron logging**

The neutron-gamma logging method records the gamma radiation emitted by nuclei in the rock when they capture thermal neutrons emitted by a fast neutron source in the probe. The neutron-capture gamma yield in a homogeneous medium is dependent on the neutron moderating properties of the medium, and is particularly sensitive to the hydrogen content in the environment. It is therefore used to identify free water in the formation or bound in the crystal lattice of rock minerals. There is also another variant of this log called the neutron-neutron log, in which the detector of gamma quanta is replaced by a proportional neutron counter. Both variants yield similar curves.

After proper calibration, the neutron log can be used for the quantitative determination of the rock porosity in the zone of saturation. Above the water table, the neutron log cannot be used for measuring the porosity, but can be very useful for measuring changes in the moisture content.

9.2.9 *Direct observation logging*

Direct observation of the physical characteristics of the strata and of the types of fissures through which water enters a well can be accomplished by either underwater photography or underwater closed-circuit television equipment. Conventional photography has the advantage that colour film can be employed, but suffers from the disadvantages that the film must be developed before results can be seen and not until this has been done will the investigators know whether or not the feature they wished to film has been recorded.

Closed-circuit television can now be employed to provide either axial views of the borehole or radial views of the borehole walls. Detailed investigations of fissure types and their geological significance are now being made using this technique. Constructional details, such as leaks in casing, can also be observed directly.

9.2.10 *Combined logging in hydrogeology, examples of interpretation*

In most cases, the hydrogeologist can get much information from uncased boreholes by combining the following logs: spontaneous potential, resistivity with short and long normal devices, and gamma ray. These logs are relatively easy to obtain.

Figure 9.2.10a shows theoretical SP, resistivity, and gamma-ray logs for common rock types and for absorbed water which is either more resistive or more conductive than the

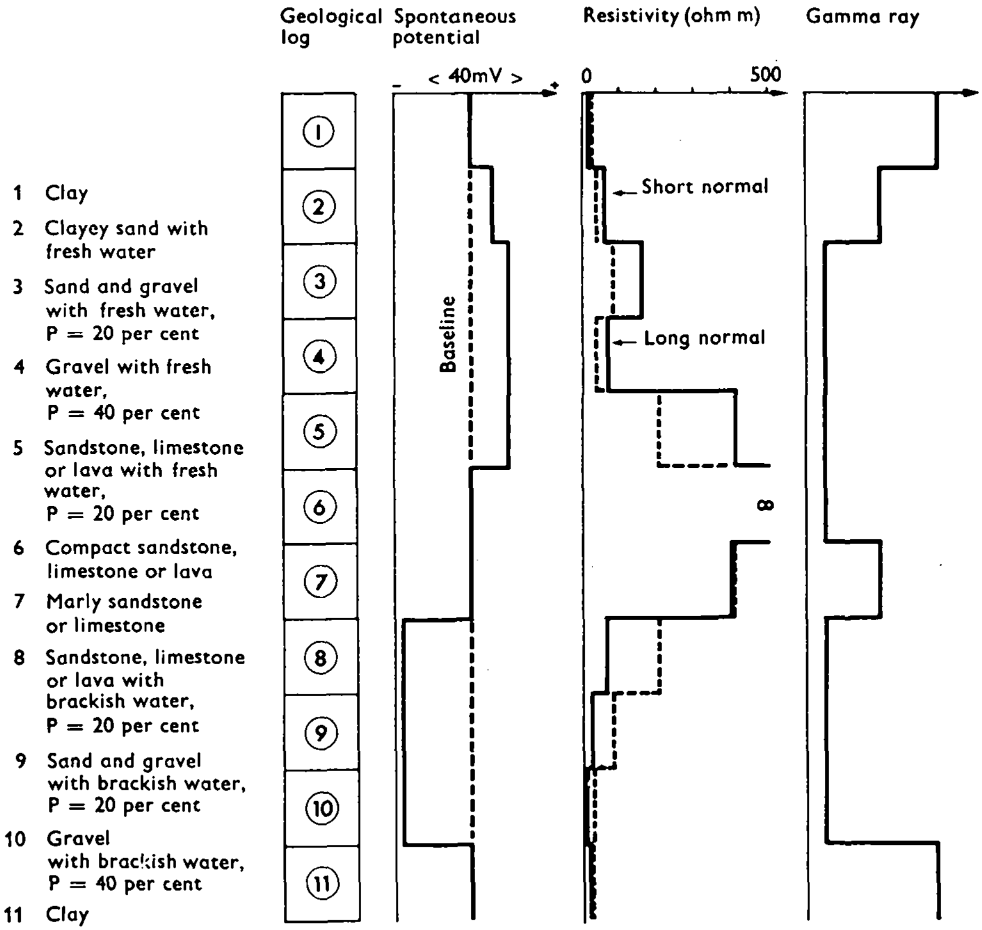


FIG. 9.2.10a. Theoretical logs: spontaneous potential, resistivity, gamma ray (calculated by J. L. Astier). Thick layers. Radius of invaded zone = radius of investigation, short normal device. Mud resistivity: 10 ohm m. Brackish water resistivity: 3 ohm m. Fresh-water resistivity: 20 ohm m. Both the mud and the ground water are supposed to be NaCl solutions. P = total porosity.

mud. In a combined analysis of these logs, the permeable layers with fresh water must satisfy three criteria simultaneously: (a) low radioactivity, indicating the absence of clay; (b) positive spontaneous potential, especially if the mud resistivity is low; (c) higher resistivity recorded with the long normal device than with the short normal, especially if the mud resistivity is low.

If the hole is cased, only nuclear logging gives usable results. Figure 9.2.10b illustrates the theoretical possibility of distinguishing between different types of rocks when a combination of the two nuclear logs is used.

Figure 9.2.10c is an example of cased holes where two nuclear logs serve as a basis for hydrogeological interpretation. At a depth of 5 10 m, an anhydrite layer can be identified by characteristic low radioactivity on the natural gamma curve and maximum radioactivity on the neutron curve indicating low porosity and lack of water. The neutron curve in borehole T-21 does not reach the maximum, probably indicating that the anhydrite was partly crystallized into gypsum. According to Keys and MacCary (1971),

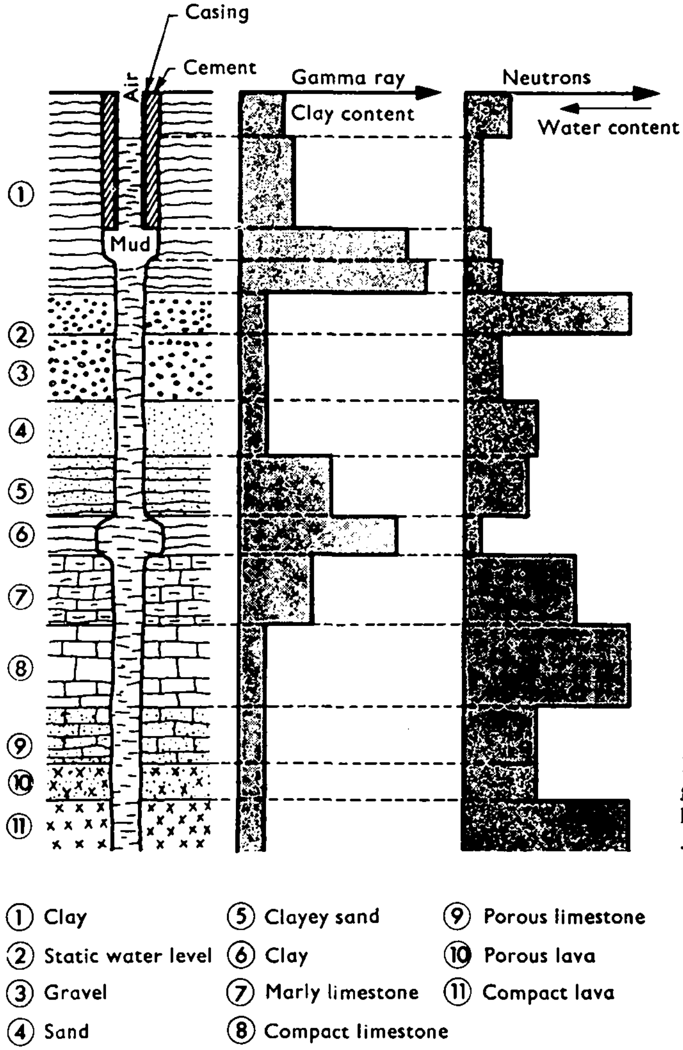


FIG. 9.2.10b. Theoretical gamma-ray and neutron logs (calculated by J. L. Astier).

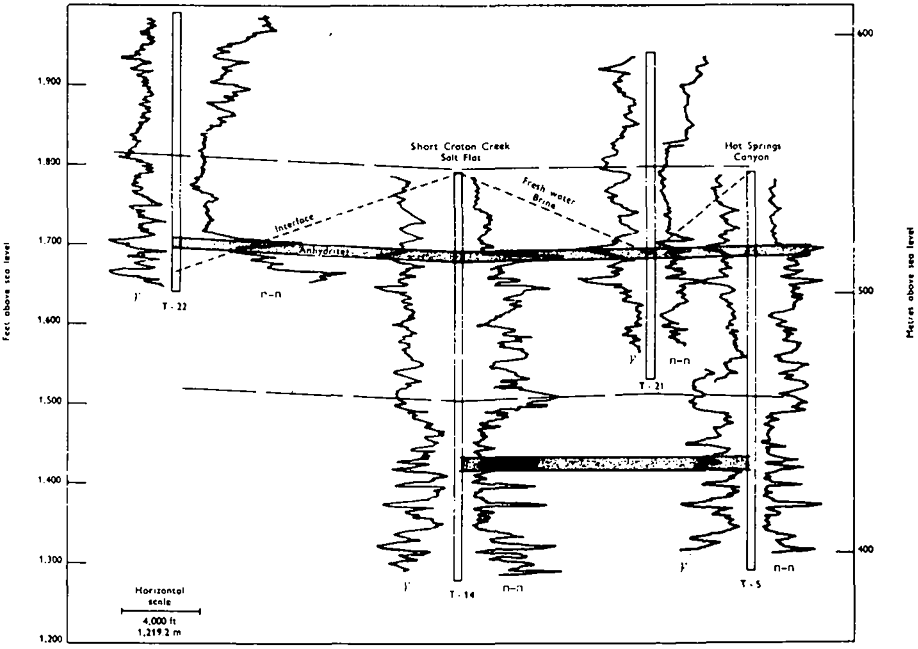


FIG. 9.2.10c. Correlation with natural gamma (γ) and neutron (n-n) logs. (Keys and MacCary, 1971.)

this indicates the depth of hydration and can be used to find the position of an interface between brine and fresh water.

The most favourable case is when the interpreter has at his disposal the results of nuclear as well as conventional logs. An example, based on real data from six different geophysical logs, is presented in Figure 9.2.10d.

The rocks named on the left side are those found in a continuous core. On the natural gamma curve below 90 m the clay content increases. This indicates that the anomaly in this sector on the neutron curve (deflection at left) may not be interpreted as an increase in the effective porosity, but is caused by the presence of water bound in the clay minerals. The zone of greatest permeability is below 30 m where there is a distinct minimum. At the same time, the gamma curve shows no clay content in this section. Anydrites, like rocks with higher density, are readily detectable on the gamma-gamma log. At the same time, the borehole diameter is enlarged by drilling through soluble halite beds and this

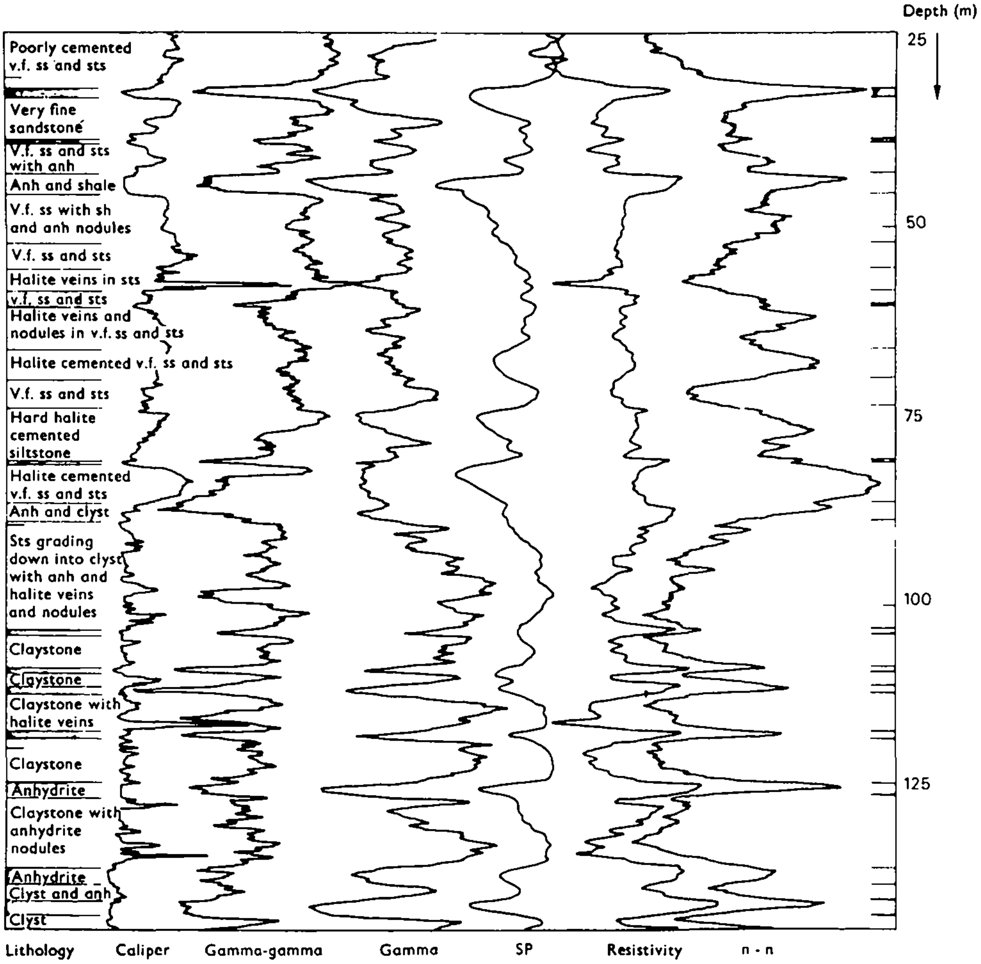


FIG. 9.2.10d. The relationship of six different geophysical logs to lithology, upper Brazos River, Texas. (Keys and MacCary, 1971.) Abbreviations: anh, anhydrite; clyst, claystone; v.f., very fine; sh, shale; ss, sandstone; sts, siltstone.

shows on the caliper log. Clay strata causing maxima on the neutron curves correlate properly with the minima on the SP line.

For more information about interpretation of results and examples of applications of nuclear logging, the reader should consult: Filippov, 1962; Ferronsky, 1969; Keys and MacCary, 1971; IAEA, 1971.

9.2.11 *Stratigraphic correlations between boreholes*

Stratigraphic correlations between boreholes are often simpler and more certain when based on geophysical logs rather than on samples or cuttings. This is especially true in sandy and clayey series. Two types of correlations are possible between logs (Fig. 9.2.11): (a) absolute correlations are made between strata for which the geophysical parameter can be clearly differentiated from the rest of the sequence because of more highly conductive clays or compact, highly resistive beds (as, for example, beds *ab* and *a'b'*; and (b) sequence correlations are based on a characteristic sequence in the geophysical parameter. The absolute values may vary from one borehole to another, but the ratios between adjacent layers vary slightly (as in the case of beds *cd* and *c'd'*, for example).

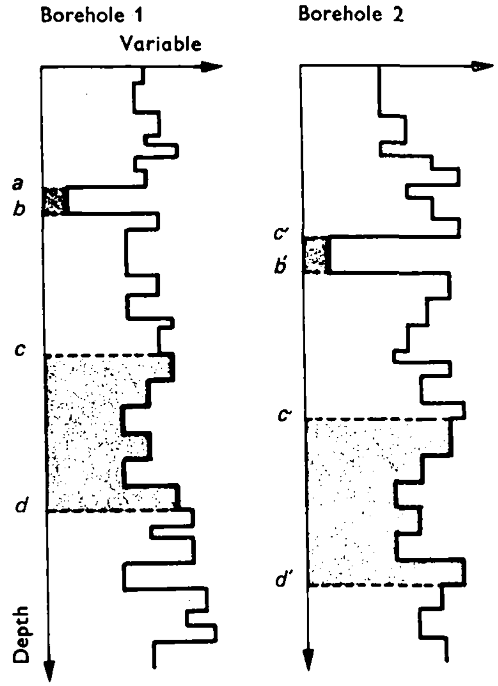


Fig. 9.2.11. Correlation between boreholes using geophysical logs: *ab*—*a'b'*, absolute correlation; *cd*—*c'd'*, sequence correlation.

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10 Nuclear techniques in ground-water hydrology[1]

10.1 Introduction

Isotope hydrology is a relatively recent field of scientific investigation, dealing with the application of isotope techniques to hydrological studies. In the last twenty years, and in particular in the last decade, a number of important results have been achieved in several types of hydrological problems by using isotope techniques. The interest in this field of studies is continuously growing, as indicated by the increasing number of laboratories and scientists which are applying isotope techniques to hydrology and by the rapid expansion of literature on the subject.

Among the many questions which may be asked of hydrogeologists about a given ground-water supply, in general the most important and critical concerns the safe yield, or for a source to be 'mined' the total yield. In order to give a satisfactory answer, many associated extremely important (and often extremely difficult) questions should be answered. Isotope techniques may solve problems such as: identification of the origin of ground water, determination of its age, flow velocity and direction, interrelations between surface waters and ground waters, possible interconnexions between different aquifers, local porosity, transmissivity and dispersivity of an aquifer. The cost of such investigations is often relatively small in comparison with the cost of classical hydrological techniques, and in addition they are able to provide information which sometimes cannot be obtained with other techniques.

Isotope hydrology can be divided into two main branches: environmental isotope hydrology and artificial isotope hydrology. The first branch uses isotope variations established in waters by natural processes. This isotopic tracing cannot be controlled by man, it can only be observed and interpreted to solve some hydrological problems on the basis of the general knowledge of isotope variations in nature. Regional hydrological problems can be studied with environmental isotopes if (as is usually the case) natural conditions establish measurable variations in isotope concentrations of different waters.

Artificial isotope hydrology makes use of radioactive isotopes artificially injected at a well-defined point of the system under investigation. Afterwards, the evolution of the isotope concentration with time is followed. The answer provided by this technique is very detailed, although its validity is limited to a restricted area around the injection point and to the conditions present at the moment of injection. Nevertheless, measurements performed at a sufficient number of points and repeated at different times may provide a good description of the hydrological system investigated.

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[1] This chapter has been prepared by the staff of the Section of Isotope Hydrology of the International Atomic Energy Agency, Vienna, Austria: E. Bradley, R. M. Brown, R. Gonfiantini, B. R. Payne (Head), K. Przewlocki, G. Sauzay, C. K. Yen, Y. Yurtsever, and revised by the Working Group on Ground-water Studies of the Co-ordinating Council of the IHD at its second session in September 1972.

In this chapter, the presentation and discussion of the basic principles of isotope techniques is limited to those which are applied to ground-water studies. More details and examples can be found in the books cited at the end of the chapter.

Advice on the suitability of isotope techniques for specific hydrological studies is available from the Section of Isotope Hydrology, International Atomic Energy Agency, Vienna (Austria). This group has accumulated considerable experience through participation in UNDP-sponsored investigations in various countries. Equipment and experts are provided in this field within the technical assistance programmes of the IAEA and the UNDP.

10.2 Environmental isotope hydrology

The isotope ratios of natural compounds may change as a consequence of the history and processes to which such compounds have been submitted in the environment. The stable environmental isotopes which are used for hydrological purposes, hydrogen-1 and -2 (deuterium), carbon-12 and -13, oxygen-16 and -18, are naturally occurring. The radioactive environmental isotopes of interest, tritium (hydrogen-3) and carbon-14, are naturally occurring, produced by interaction of cosmic radiation with the atmosphere, and man-made in the explosion of nuclear devices. The man-made isotopes have been dispersed throughout the earth's atmosphere and their participation in the hydrological cycle is now regulated by natural processes.

The chemical and physical techniques and equipment which are required for measurement of environmental isotope variations with sufficient accuracy were developed in the early 1950s. In the next few years, investigations established the general scientific background for the understanding and interpretation of isotope variations in natural compounds. Towards the end of the fifties, the first papers dealing with the application of environmental isotopes to hydrological problems appeared in the scientific literature.

Isotopes of oxygen and hydrogen, the chemical elements which constitute the water molecule are, in a certain sense, ideal geochemical tracers of water because their concentrations are not subject to changes by interaction with the aquifer material. On the other hand, carbon compounds in ground water may interact with the aquifer material, complicating the interpretation of ^{14}C data.

A few other environmental isotopes such as ^{32}Si (Lal *et al.*, 1970) and $^{238}U/^{234}U$ (Kaufman *et al.*, 1969) have been recently proposed, but their use has been quite limited until now and they will not be discussed in this chapter.

10.2.1 *Stable isotopes of hydrogen and oxygen in the hydrological cycle*

The stable isotopes of hydrogen and oxygen, the two elemental constituents of water, and their mean abundances in natural compounds are reported in Table 10.2.1.

TABLE 10.2.1. Main environmental isotopes used in hydrology

Isotope	Relative abundance in nature (%)	Decay	Half-life	Maximum energy (KeV)
^{1}H	99.985	Stable		
$^{2}H = D$	0.015	Stable		
$^{3}H = T$	$10^{-15} - 10^{-12}$ *	β^-	12.26 y	18.1
^{12}C	98.89	Stable		
^{13}C	1.11	Stable		
^{14}C	1.2×10^{-10} **	β^-	5,730 y	156
^{16}O	99.76	Stable		
^{17}O	0.04	Stable		
^{18}O	0.20	Stable		

* The lower figure refers to the tritium content of precipitation before 1952. The higher to that reached in 1963 precipitation in the northern hemisphere.
** In modern carbon before 1950.

Among all isotopic species of water possible, only these are of practical interest: $H_2{}^{16}O$, $HD^{16}O$, $H_2{}^{18}O$. The variations of the isotopic ratios D/H and $^{18}O/^{16}O$ in water samples are expressed in terms of per mille difference (δ‰) with respect to the isotopic ratios of mean ocean water, which constitutes the reference standard SMOW (Craig, 1961*b*):

$$\delta‰ = \left(\frac{R_{\text{sample}}}{R_{\text{SMOW}}} - 1\right) \times 1{,}000$$

where R is the isotope ratio D/H or $^{18}O/^{16}O$. Thus a sample which has $\delta^{18}O = +5$ (or -5) has an ^{18}O content 5‰ higher (or lower) than that of the mean ocean water. The same applies to the δD for the deuterium content.

Using the standard techniques of sample preparation described in the early papers (Friedman, 1953; Epstein and Mayeda, 1953) and the Nier-McKinney mass spectrometer (McKinney *et al.*, 1950) or one of its derivatives now commercially available, standard deviations of 1.0‰ for δD and 0.1‰ for $\delta^{18}O$ can be achieved. These deviations, for practical purposes, are comparable because the D/H variations in natural waters are, in most cases, 5 to 8 times larger than those of $^{18}O/^{16}O$. This measurement accuracy is sufficient for hydrological studies.

10.2.1.1 Isotopic variations in precipitation

The mass difference between isotopes of the same element produces small differences in their chemical and physical behaviour (Bigeleisen, 1965), which in turn establish slightly different isotopic compositions (isotope fractionations), among co-existing chemical compounds or phases.

Many natural processes cause variations of the isotopic composition of natural waters. Among them, the most important are evaporation and condensation. During the evaporation process, the light molecules of water, $H_2{}^{16}O$, are more volatile than those containing a heavy isotope, i.e. $HD^{16}O$ or $H_2{}^{18}O$. Therefore, water vapour which evaporates from the ocean is depleted by about 12-15‰ in ^{18}O and 80-120‰ in deuterium with respect to the ocean water. When this atmospheric water vapour undergoes successive cooling and condensation with production of clouds and precipitation, the less volatile heavy molecules condense preferentially, leaving a residual vapour more and more depleted in D and ^{18}O. As a result, successive precipitations derived from the same initial vapour mass will be more and more depleted in heavy isotopes. Because the degree of condensation of a vapour mass depends on the temperature, a relation between isotopic composition of precipitation and its temperature of formation should be expected: as the formation temperature decreases, the δ-values of precipitation decrease (Dansgaard, 1954). This has been observed directly in Antarctic precipitation (Picciotto *et al.*, 1960) and a world-wide relation between $\delta^{18}O$ of precipitation and mean annual air temperature has been reported (Fig. 10.2.1.1a, Dansgaard, 1964).

This dependency on temperature produces seasonal isotope variations of precipitation (winter precipitation is depleted in heavy isotopes with respect to summer precipitation), latitude variations (high latitude precipitation is depleted with respect to low latitude precipitation) and altitude variations (the heavy isotope content of precipitation decreases with increasing altitude); (Friedman *et al.*, 1964; Moser and Stichler, 1970). The last effect is especially important in regional hydrological studies, where for instance ground waters deriving from recharge areas at different elevations may be differentiated. The altitude effect may change from region to region but, in general, it is about 0.3‰ decrease in ^{18}O content and 2.5‰ decrease for D content per 100 metres increase in elevation.

The variations of the deuterium and oxygen-18 contents in precipitation are linearly correlated (Fig. 10.2.1.1b, Craig, 1961*a*). The relationship:

$$\delta D‰ = 8\delta^{18}O‰ + 10 \qquad 10.2(1)$$

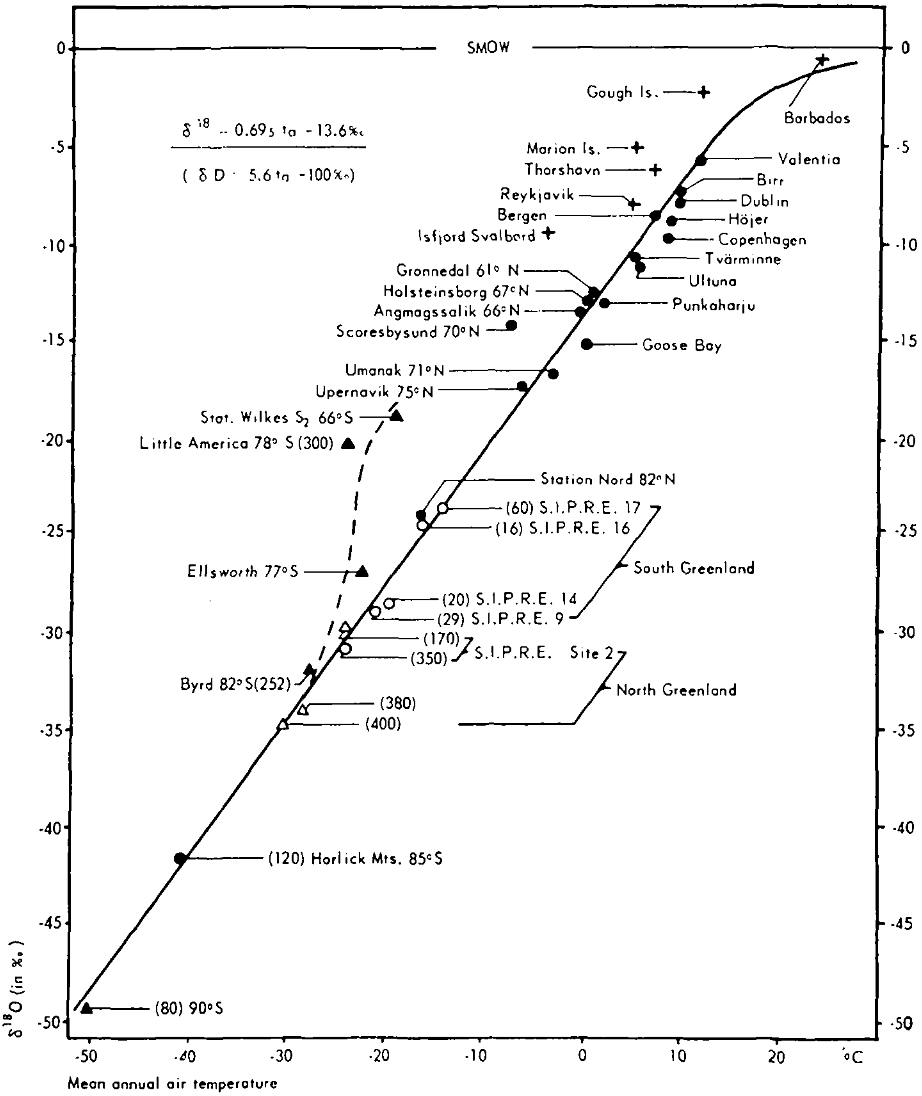

FIG. 10.2.1.1a. The annual mean $\delta^{18}O$ of precipitation as a function of the annual mean air temperature at surface. The figures in parentheses indicate the total thickness (in cm) of the investigated snow layers (Dansgaard, 1964).

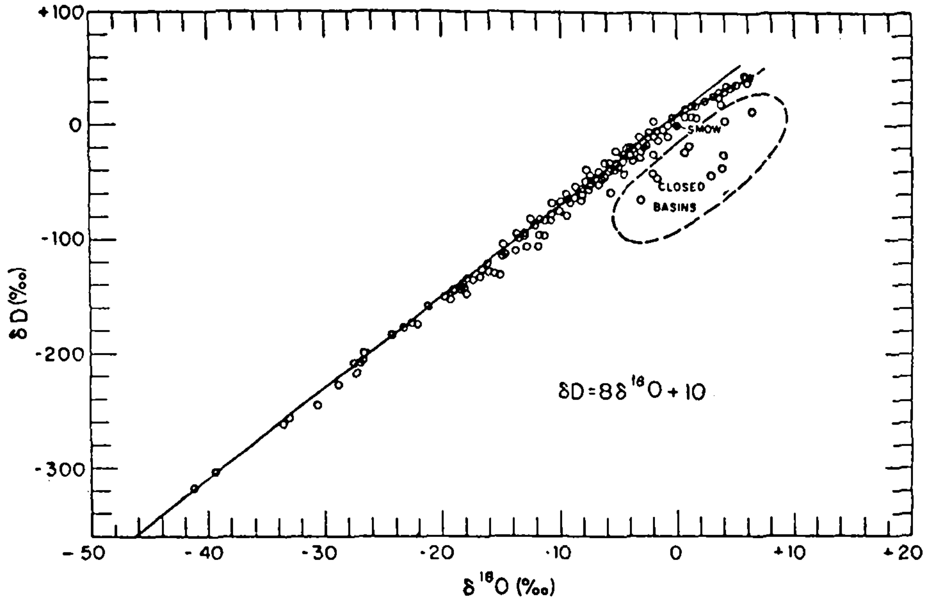

FIG. 10.2.1.1b. Deuterium and oxygen-18 correlation in precipitation and surface waters. Closed basin waters do not fit the general linear correlation, because of evaporation. Points which fit the dashed line at the upper end of the curve are rivers and lakes from East Africa (Craig, 1961*a*).

best fits the points representing the isotopic compositions of precipitation samples from all over the world, at least for values of $\delta^{18}O$ lower than zero with respect to SMOW. In some regions of the world, a different intercept on the δD axis is observed, but a slope of 8 is generally preserved.

Precipitation which has undergone significant evaporation during its fall does not obey Equation 10.2(1). Evaporation does tend to enrich both the heavy isotopes in water but not in the relative proportion indicated by the above relationship (Craig, 1961*a*; Ehhalt *et al.*, 1963; Woodcock and Friedman, 1963).

10.2.1.2 **Isotopic composition of ground water**

When precipitation infiltrates to feed ground water, mixing in the unsaturated zone smoothes the isotopic variations so that water in the saturated zone has a composition corresponding to the mean isotopic composition of infiltration in the area. This may differ slightly from the mean isotopic composition of precipitation in the area due to the fact that not all the precipitation during the year infiltrates in the same proportion (Thatcher, 1967; Gat and Tzur, 1967). For example, at temperate continental sites spring and summer precipitation partially or totally re-evaporates from the soil before infiltration can occur.

In the aquifer, the isotopic composition of water does not change further unless exchange with the oxygen of the rocks occurs. This process of exchange is, in general very slow at the temperatures normally occurring in aquifers, and is of importance only for high temperature thermal waters (Craig, 1963).

The isotopic composition of ground water is thus related to that of precipitation in the recharge region of the aquifer at the time of recharge. Ground water may be of a very old age, of the order of several thousands or tens of thousands years, and the climatic conditions of the region at the time of recharge may have been different from today. This implies that the isotopic composition of precipitation could have been different from the present one, due to the correlation between δ values and temperature (Gat, 1971).

Ground water may be recharged also by lateral seepage from surface waters, such as river and lake waters, or by vertical infiltration from permanently or temporarily ponded waters. If most of the recharge is from lateral seepage, the ground water should reflect the mean isotopic composition of the river or the lake instead of that of local precipitation which could be rather different. The river may collect water which originates from precipitation in a completely different area, for instance in a high mountain region; in this case its heavy isotope content would be lower than that of precipitation in the plain, due to the altitude effect.

In the case of lakes or ponds, the water may be considerably enriched in heavy isotopes through evaporation. The enrichment is limited by direct isotopic exchange with atmospheric moisture. It is clear that this enrichment is higher where the evaporation is more intense with respect to the total volume of water, e.g. in closed lakes and ponds or lakes and rivers in arid areas. In the latter case, the atmospheric relative humidity is low and thus the isotopic exchange is less important.

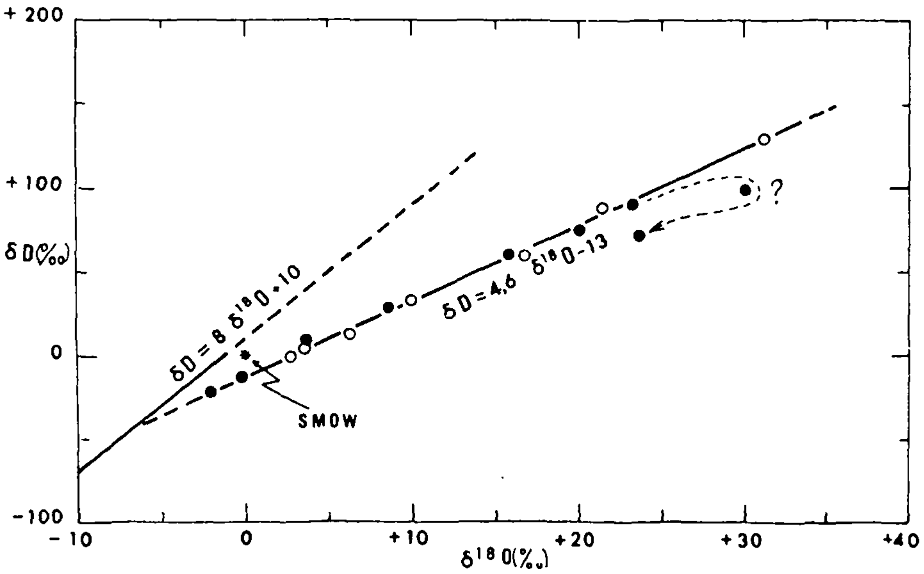

FIG. 10.2.1.2. Deuterium and oxygen-18 correlation in two Saharian lakes going dry. One of them contained salt water (filled points): at high values of the salt concentration, the isotopic composition of water deviates from the linear correlation (Fontes and Gonfiantini, 1967).

Waters which have undergone evaporation are easily recognized by their isotopic composition (Figs. 10.2.1.1b and 10.2.1.2). Their heavy isotope content is higher than that of non-evaporated waters in the region and they do not obey the relationship in Equation 10.2(1) (Craig and Gordon, 1965; Fontes and Gonfiantini, 1967; Dinçer, 1968).

10.2.2 *Tritium in the hydrological cycle*

Tritium is a radioactive isotope of hydrogen having mass 3, half-life 12.26 y, β-radiation of maximum energy 18.1 KeV, no γ radiation. It occurs in the environment as a result of both natural and man-made processes. It is produced naturally by the interaction of cosmic radiation with the nitrogen and oxygen of the upper atmosphere at a rate of about 0.25 atoms/cm^2/sec (Lal and Suess, 1968). Large amounts of man-made tritium were released to the atmosphere by thermonuclear tests in the period 1953-62, and minor amounts are being released by industrial nuclear activities.

Most tritium produced in the atmosphere is rapidly oxidized to HTO and incorporated in the hydrological cycle where it constitutes a very useful marker for water that has been in the atmosphere within the past twenty years. The great dilution by H_2O results in very low tritium concentrations which can be measured only by means of tritium's radioactivity, usually after an isotopic enrichment treatment.

The tritium content of natural waters is expressed in Tritium Units (T.U.). One Tritium Unit corresponds to a concentration of 1 tritium atom per 10^{18} hydrogen atoms. This is about the lower limit of detection for most laboratories.

10.2.2.1 **History of tritium in the atmosphere**

Cosmic radiation establishes a concentration of about 10 T.U. in temperate zone continental meteoric waters. This was the concentration observed before 1953 in northern hemisphere precipitation. After 1953, the tritium content of precipitation increased as a result of thermonuclear testing (Brown and Grummitt, 1956; Brown, 1961). Values up to 10,000 T.U. were reached in the northern hemisphere in 1963 following extensive testing in 1961-62 (Thatcher and Payne, 1965). From 1963, the tritium content of precipitation decreased until 1968, as a consequence of the moratorium established for the explosion of thermonuclear devices in the atmosphere. Since 1968, the tritium decrease has considerably lessened, or even stopped. The history of tritium in precipitation is reported in Fig. 10.2.2.1a.

Most thermonuclear tritium was deposited in the stratosphere. About half of the stratospheric inventory is transferred to the troposphere in the spring and summer months of each year (Brown, 1970) resulting in a continuing seasonal cycle of tritium precipitation, with maximum in summer and minimum in winter.

There is considerable geographical variation of the tritium content of precipitation. Lower values occur at oceanic and coastal sites than at continental sites because the ocean serves as a sink for HTO through isotopic exchange between the atmospheric water vapour and ocean water having low tritium content. Low concentrations occur in equatorial and southern regions as the result of several factors. Most tritium was released in the northern hemisphere and is transferred from the stratosphere to the troposphere preferentially at high latitudes. In addition, the higher proportion of ocean surface in the southern hemisphere and the high vapour pressure in equatorial regions provide more 'wash-out' and dilution. The tritium content of precipitation in the southern hemisphere is about one tenth that in the northern hemisphere (Fig. 10.2.2.1b).

The IAEA (1969, 1970, 1971*a*) publishes data on the concentration of tritium in precipitation collected at a large number of stations around the globe, from which it is possible to estimate the tritium deposition at most places of interest.

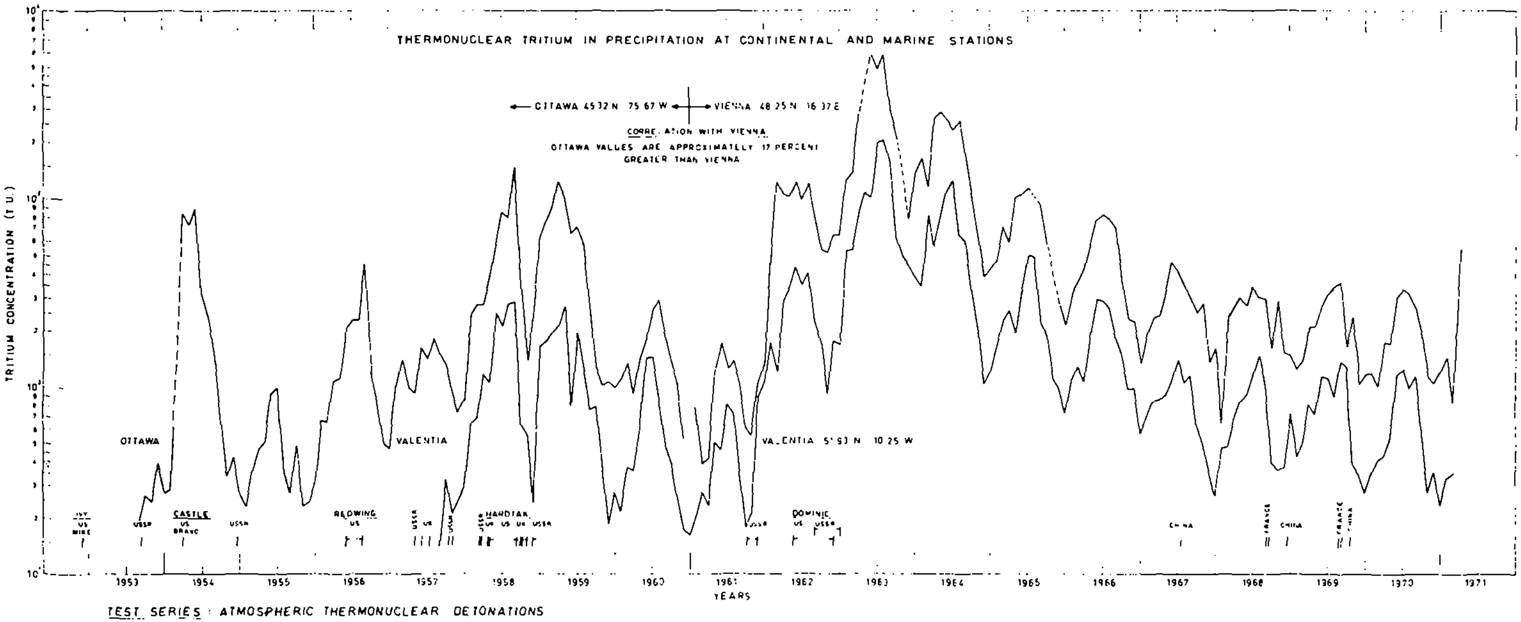

FIG. 10.2.2.1a. Tritium concentration in precipitation in Ottawa (Canada), Valentia (Ireland) and Vienna (Austria) from 1953 to 1970 (Brown, 1970; IAEA network data). The major atmospheric thermonuclear explosions are indicated at the bottom of the diagram.

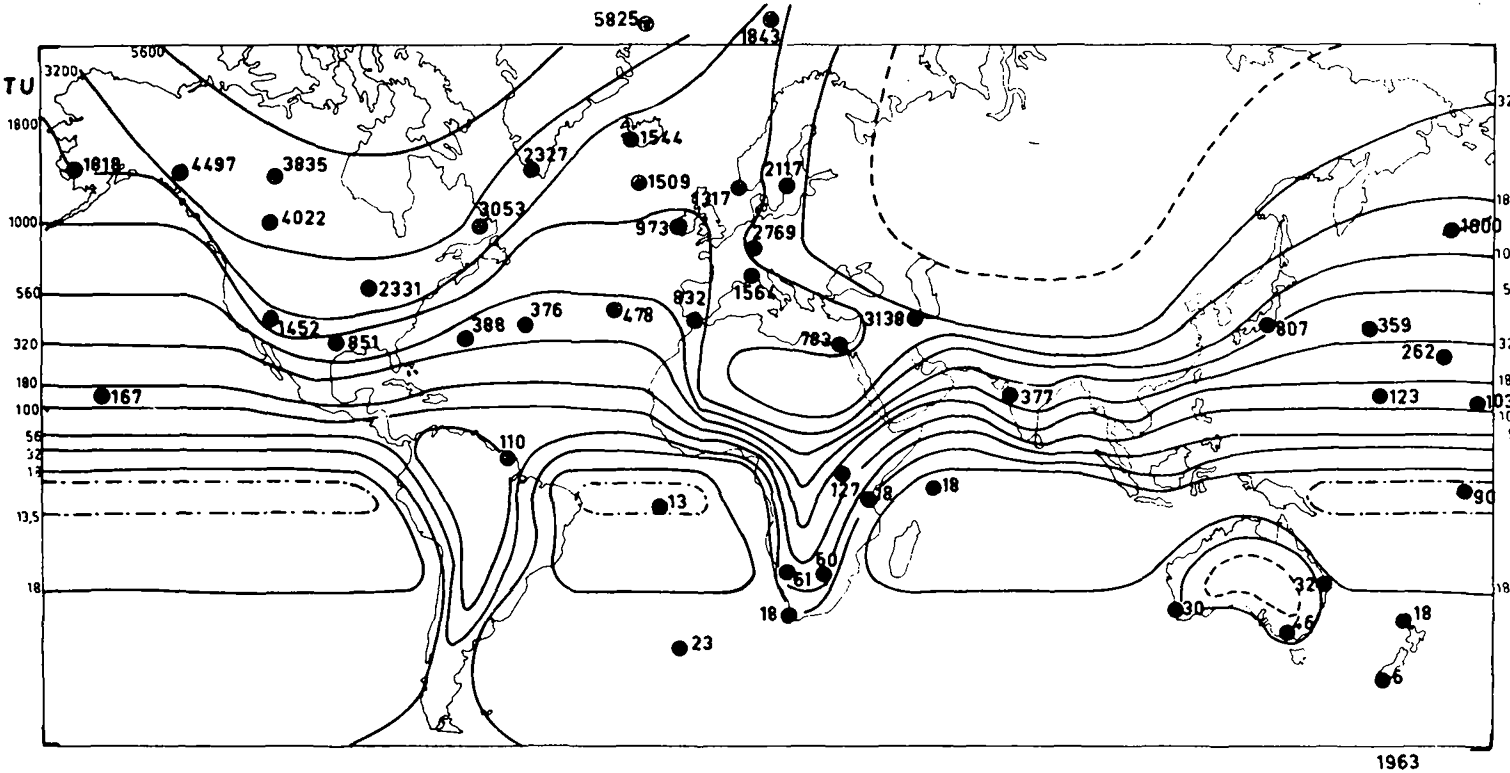

FIG. 10.2.2.1b. Geographical distribution of the tritium concentration (in Tritium Units) in precipitation in 1963 (*Guidebook on Nuclear Techniques in Hydrology*, Vienna, IAEA, 1968).

10.2.2.2 Tritium in ground water

The tritium content of precipitation is used to estimate the input of tritium to ground-water systems. Allowance must be made for the fact that recharge is often seasonally selected. In vegetated areas, most precipitation (and tritium) of the growing period may be returned to the atmosphere by evapotranspiration. At semi-arid sites, light precipitation is evaporated from the soil before it can infiltrate.

In ground-water studies, tritium measurements give information on the time of recharge to the system. In the absence of mixing, the following cases (considered for a northern hemisphere continental site in 1972) can occur:

1. *The water has a concentration of* < 3 *T.U.* This means that no water younger than 20 years is present. That is, more than 20 years are required for water to reach the sampling point from the recharge area. This is the case of most confined aquifers. Phreatic aquifers can have low tritium content due to (a) very slight infiltration (arid and semi-arid regions); (b) long percolation time (low transmissivity, great depth of water table); (c) age stratification of water below the water table.
2. *The tritium content is 3–20 T.U.* A small amount of thermonuclear tritium is present, indicating most probably water of the first test period, 1954-61.
3. *The tritium content is* > 20 *T.U.* The water of high tritium content is obviously of recent origin. If variations occur through the year and are related to the variations in precipitation over the recharge area, the flow-through is rapid and direct and the transit time may be evaluated from the time-lag in appearance of the annual peaks. The variability may also be caused by a seasonal change in the source of the water or the relative proportions of water from different sources, e.g. a tritium-free water of deep circulation and a young water of high tritium content, generally of a more superficial circulation.

The foregoing implies that ground water remains quite discretely segregated according to age during transit. In practice, this is not the case because of dispersive effects of the granular aquifer material. Tritium peaks and valleys are smoothed out and the above guidelines indicate periods rather than specific years of recharge. All cases may be subject to mixing of waters of different ages having widely different tritium content. A small portion of recent water combined with a major portion of tritium-free water may look like cases 1 or 2 above.

A further condition is often observed: *Tritium content of 20-200 T.U. showing no fluctuation* and changing only slowly. This indicates young water well mixed in the aquifer with older water. The size of the ground-water reservoir dampens all the fluctuations in the recharge. The system may be treated as a 'well-mixed reservoir' showing exponential discharge and a Mean Residence Time of water within the system estimated (Nir, 1964; Dinçer and Davis, 1967).

10.2.2.3 Infiltration studies

Tritium has been used to study the downward progress of water in the unsaturated zone, the usual means of recharge to a ground-water system. Water artificially enriched in tritium and deuterium was used to mark a specific horizon of soil moisture. Its rate of downward movement and the dispersion of the tracer were observed in core samples subsequently taken from the test plot (Fig. 10.2.2.3, Zimmermann *et al.*, 1967). The downward progress of environmental tritium shows the infiltration of many years time in a completely undisturbed system (Anderson, 1965; Smith *et al.*, 1970).

10.2.3 *Carbon isotopes in the hydrological cycle*

Carbon-14 emits beta radiation with a maximum energy of 156 KeV and has a half-life of 5,730 y. Like tritium, it occurs as the result of both natural and man-made processes.

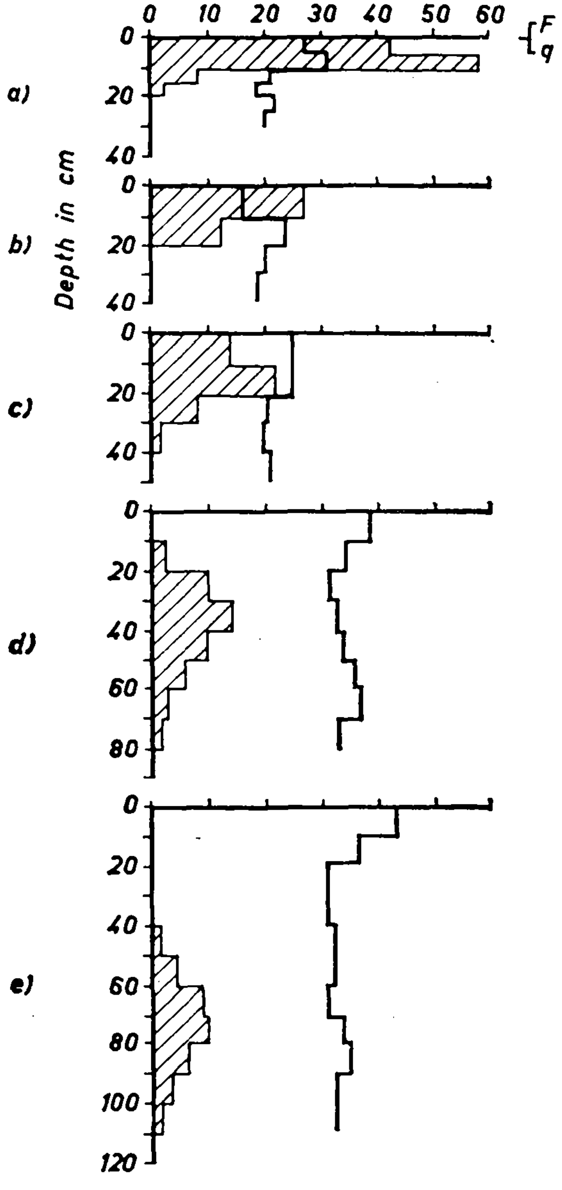

FIG. 10.2.2.3. Tritium content and soil moisture profiles at different times in a field experiment in Giessen (Federal Republic of Germany). F (heavy line) soil moisture in volume per cent; q (boundary of hatched area), amount of tritium per 10 cm layer of soil as a percentage of the amount of tritium originally deposited. Loamy soil without vegetation, labelled 9 July 1964. (a) to (e), samplings 4, 26, 82, 152 and 214 days later, respectively (Zimmermann *et al.*, 1967).

The natural production by the interaction of cosmic ray-produced neutrons with ^{14}N atoms in the total atmosphere is estimated to be 2.5 atoms/cm^2/s. (Lal and Suess, 1968). The carbon-14 is then oxidized to carbon dioxide, mixes with the carbon dioxide of the atmosphere and enters the global carbon cycle.

The natural carbon-14 concentration has been demonstrated to vary during the last seven thousand years or so by measurements of the ^{14}C content of tree rings of the *Sequoia gigantea* and Bristlecone pine (Damon *et al.*, 1966; Suess, 1967). However, the uncertainty this causes in using ^{14}C in hydrology is much smaller than other errors which may arise in this application.

The ^{14}C concentration is expressed as a percentage of the modern (pre-bomb) ^{14}C content of plants. Production of ^{14}C by the detonation of nuclear explosives caused the atmospheric ^{14}C concentration in the northern hemisphere to about double by 1963. Since then it has fallen but still remains above the pre-bomb level. Increases were also observed in the southern hemisphere, but the peak value was only about 60 per cent above the pre-bomb level. Clearly these variations are only of importance when young waters are being measured for their ^{14}C content (Olsson, 1968).

10.2.3.1 Basis of carbon-14 dating of ground water

The use of ^{14}C for dating of ground water was first proposed fifteen years ago (Münnich, 1957). It is based on the fact that soil-zone carbon dioxide is of biogenic origin, resulting from the respiration of plant roots and plant decay and hence contains ^{14}C derived by plants from the atmosphere. This biogenic CO_2 dissolves in infiltrating water and is carried down to the ground-water reservoir. Its ^{14}C content decreases by radioactive decay and the fraction of the original remaining is a measure of the time since it was removed from the soil zone, i.e. the time since infiltration of the associated water. The equation which is applied is:

$$t \text{ (years)} = 8{,}270 \ln \frac{C_0}{C} \qquad 10.2(2)$$

where 8,270 is the mean life of ^{14}C in years, C_0 is the initial ^{14}C concentration and C is the ^{14}C concentration in the sample.

The ^{14}C is measured relative to the total carbon content of the sample, so one must consider the origin of both the ^{14}C and the stable carbon of the sample. Not all of the stable carbon of ground-water carbonate is of the same origin as the ^{14}C. Infiltrating water, containing carbon dioxide dissolved from the soil zone, dissolves carbonate minerals in the soil:

$$\begin{aligned} H_2CO_3 &\rightleftarrows HCO_3^- + H^+ \\ CaCO_3 + H^+ &\rightleftarrows HCO_3^- + Ca^{++} \end{aligned} \qquad 10.2(3)$$

However, the carbon originating from limestone in general contains no ^{14}C, so that the water reaching the water table contains dissolved carbon (in the chemical forms of H_2CO_3, HCO_3^- and CO_3^{--}) with a ^{14}C content lower than that present in the soil biogenic CO_2. The evaluation of the dilution of soil CO_2 originally containing 100 per cent of modern ^{14}C with ^{14}C-free carbonate to estimate the initial ^{14}C concentration in recharge water reaching the water table constitutes the most difficult problem of the ^{14}C age determination of water.

10.2.3.2 Correcting for mineral carbonate

Different procedures have been proposed for the evaluation of the initial ^{14}C concentration.

On the basis of a large number of ground-water ^{14}C analyses from Europe and Africa, it appears that only a few samples show more than 90 per cent (relative to modern) ^{14}C content, and some of them are clearly contaminated by thermonuclear ^{14}C. There is, on the contrary, a large number of samples which contain 80 to 90 per cent of ^{14}C: therefore, a value of 85 per cent would be a good average for the ^{14}C content of recent, prethermonuclear waters. This value has been proposed as the initial ^{14}C content of carbon dissolved in ground water (Vogel, 1970).

Another method for evaluating the initial ^{14}C content of ground water uses the different $^{13}C/^{12}C$ ratios of biogenic CO_2 and limestone. (Mean abundance $^{12}C = 98.9$ per cent, $^{13}C = 1.1$ per cent). The ^{13}C content is measured by mass spectrometry and expressed as per mille relative deviation ($\delta^{13}C‰$) from the standard PDB (Craig, 1957), which has a $^{13}C/^{12}C$ ratio close to that of marine limestone. In temperate climates, soil biogenic CO_2, derived from plant respiration and the decay of organic matter, has a $\delta^{13}C$ value of $-25 \pm 3‰$, corresponding to that observed in local plants. Isotopic fractionation on dissolution of this CO_2 in water and conversion to bicarbonate raises the value to $-17 \pm 3‰$. Limestone has, in general, a $\delta^{13}C$ value of $0 \pm 2‰$. The $\delta^{13}C$ values of carbon species dissolved in ground water range from -20 to -5 in most cases. A simple proportion gives the fraction of biogenic carbon present in a ground water which corresponds to the fraction of modern ^{14}C present in ground water at the time of recharge (Pearson, 1965).

The foregoing treatment is in many cases inadequate because it does not take into account the over-all chemistry of the system (Pearson and Hanshaw, 1970). For instance, knowledge of the pH is necessary to evaluate the relative amounts of H_2CO_3 and HCO_3^- present and to take into account the isotopic fractionation occurring between them.

The effect of other ions, such as sodium and magnesium, on the carbonate equilibrium cannot be ignored. In the case of sodium bicarbonate waters, calcium ions may have been removed from solution by exchange with the sodium of clay minerals, so permitting additional $CaCO_3$ to be dissolved. This would then result in more positive $\delta^{13}C$ values.

Another difficulty, especially in arid areas or areas where important climatic changes may have occurred since the time of infiltration of water, is our limited knowledge of the climatic condition effect on the isotopic composition of vegetation (and then of the soil CO_2). At present, there are relatively few published data for plants in arid regions, but $\delta^{13}C$ values of about $-12‰$ are not uncommon. This may affect strongly the evaluation of the initial ^{14}C content based on ^{13}C values.

The possibility of exchange of ^{14}C between the dissolved bicarbonate and the carbonates of the matrix of the aquifer material has been recognized (Thilo and Münnich, 1970). However, it appears that this effect is small in nature for waters which have not been submitted to high temperatures. For thermal waters there is some evidence for exchange which, therefore, requires that interpretation of ^{14}C data be treated with caution in such cases.

10.2.3.3 **Applicability**

The ^{14}C method can be used for waters younger than 30,000 years. In general, it is applied to study the movement of water in confined aquifers. Where recharge occurs only in the outcrop area and the water chemistry and isotopic composition of dissolved carbon species are relatively uniform, the age differences in space are not affected by the uncertainties which affect the absolute age determination of water. Thus, it is possible to determine the flow velocity of water by determining the age differences between two sampling points at known distance. This provides the hydrologist with an estimate of the mean regional permeability.

Carbon-14 measurements may also give information on mixing processes of waters of different ages within a given aquifer.

10.2.4 *Project planning*

Isotope techniques are a supplementary tool for hydrogeological investigations. The amount and the quality of information that they are able to provide is strictly dependent on the degree of hydrogeological knowledge of the area under investigation. It is evident, for instance, that selection of the sampling points in a given area must be made on the basis of the information available for each point. Interpretation of the results will be more reliable when the main hydrogeological features of the ground-water body being studied are well known. Environmental isotope techniques should not only be applied with conventional hydrological methods, but usually a fuller interpretation is possible if capability for all environmental isotopes is available. In this way there is a valuable interplay between the different data which leads to the optimum use of these techniques.

10.2.4.1 **Development of an isotope hydrology study**

Ideally, an isotope hydrology study should develop as follows. The field hydrogeologist first makes the best possible appraisal of the area under investigation, defines the main hydrological problems and establishes possible working hypotheses. At this point, the hydrogeologist and the expert in the application of isotope techniques in hydrology (whom we may refer to as an isotope hydrologist) select and discuss those problems which could be studied with isotope techniques, and the hypotheses which could be tested. Although a visit of the area by the isotope hydrologist may not seem strictly necessary, in practice it is highly recommended. The more realistic picture gained of the area under investigation is almost essential for correct selection of sampling points and subsequent interpretation of results. The visit also provides the isotope hydrologist with an opportunity to instruct the local personnel, under field conditions, on sampling techniques required to avoid contamination or alteration of samples. Sometimes, the isotope hydrologist will recognize unforeseen site conditions that necessitate special precautions which would not seem relevant to a person who is not familiar with isotope techniques.

At least two sampling campaigns are recommended whenever possible. The samples collected in the first campaign should cover all the main problems under investigation, but not necessarily all of them will be isotopically analysed. The results obtained from this first set of reconnaissance samples will provide answers to basic questions such as: Are there significant isotopic differences among the various waters in the system which may serve to identify them? What are the ranges of values for tritium and ^{14}C? How does this isotopic data agree with the hydrological hypotheses? On the basis of this information, a second more detailed sampling should be organized for those problems for which isotopic techniques look promising.

In the case that variations of isotopic composition with time are expected (for instance in karstic systems, or any system having rapid circulation) sampling should be carried out regularly in time (say, two to four samples per year at each sampling point). If recharge is likely to be markedly seasonal, e.g. because of precipitation distribution or spring melt, it is advisable to schedule samplings for the end of the wet and dry, or flood and base flow periods. In the first stage of the study only a few samples need be analysed to learn of the degree of variability occurring. On the basis of the results, it should be decided whether it is worth while to make more analyses.

The interpretation of isotopic results and the preparation of the report should be made by collaboration of the isotope hydrologist and the hydrogeologist. Some hypotheses could seem reasonable from the hydrological standpoint, but be unacceptable from the point of view of the isotopic results, and vice versa. It is then necessary to decide which of the possible hypotheses best fits all the isotopic, geochemical, geophysical and hydrogeological data.

10.2.4 2 **Sampling**

In studying the isotope content of ground water, a major requirement is to obtain an uncontaminated sample representative of a given aquifer or zone. This requires careful selection, if feasible, of wells for which relevant information is available. A few carefully selected samples are more valuable than dozens for which the necessary supporting data are unreliable.

It is important to avoid contamination of the sample from upper zones penetrated by the well. This is especially important in tritium and carbon-14 sampling because in many areas the uppermost ground water may have a much higher concentration of these isotopes than the zone of interest. Sufficient water must be removed by pumping or other means to replace the volume of water standing in the well before a sample is taken for isotope analysis. Water standing in an unused well will generally give spurious isotope results.

Sampling of wells during drilling offers possibilities of determining isotopic stratification of water, but contamination of samples with drilling fluid may easily occur. Samples should be taken only after sufficient water has been removed from the hole to assure that formation water has been sampled.

For stable isotopes and tritium, relatively small samples are required: 20 ml are sufficient for stable isotopes, 500 ml for tritium. Containers must be airtight to avoid evaporation and exchange. Plastic bottles with inner polyethylene cones in their screw caps are generally used. Bottles should be filled completely and care taken to minimize exposure of the samples to the atmosphere. For long-term storage (more than 1 year) samples should be kept in well-sealed glass bottles.

Carbon-14 sampling is done by separating dissolved carbonate and CO_2 from a large enough volume of water (usually 50-100 l) to yield 3-4 g of carbon, preferably at the sample site. Two different methods may be used: precipitation or gas evolution. In the first method, the water sample is adjusted to pH ≈ 9 with carbonate-free NaOH solution and $BaCl_2$ solution added to precipitate $BaCO_3$. The $BaCO_3$ precipitate is allowed to settle into a bottle for shipment to the laboratory for ^{14}C analysis (Fig. 10.2.4.2). A detailed description of this method is distributed on request by the Section of Isotope Hydrology of IAEA.

The second method consists of acidifying the water sample to release its CO_2 which is purged by a circulating gas system and collected in an alkaline solution for shipment to the laboratory (Vogel, 1967).

In both methods, it is important to minimize exposure of the initial water sample and the carbonate concentrate to the air to avoid pick-up of atmospheric CO_2 having modern carbon-14.

Information to accompany the samples includes data normally part of a well or spring history, such as: place of sampling, date, the depth of the well, interval tapped, aquifer, static level, water temperature, pH, chemical composition, depth of water, whether aquifer is confined, etc.

Sample transport and storage are a small part of the expenses of a field trip and subsequent analysis, so it is desirable to take more samples during the initial stages of the investigation than might at first seem necessary. Sampling cannot be backdated!

10.2.4.3 Costs of analyses

A number of commercial firms and university laboratories in various countries perform environmental isotope analyses. The IAEA Isotope-Hydrology Laboratory does analyses specifically for United Nations-sponsored projects. Costs of analysis are dependent on the number of samples and, in the case of tritium, on the concentration range. Current price ranges (1972) are: ^{13}C, D or ^{18}O, \$15-25; tritium, \$50-100; carbon-14, \$80-150.

Equipment for the measurement of tritium and carbon-14 costs about \$30,000. A mass spectrometer for stable isotope measurements costs about the same. Specially trained personnel are required for the successful operation of such equipment.

FIG. 10.2.4.2. Carbon 14 sampling in the field. The container (about 70 l) is filled with water, and the $BaCO_3$ precipitate is collected in the bottle at bottom. [Photo: PAHO].

10.2.5 *Examples of application*

10.2.5.1 Interconnexions between ground water and lakes

Oxygen-18 was used to test a hypothesis that Lake Bracciano (in Central Italy) was leaking into a near-by phreatic aquifer, the water table of which was 15 to 20 m lower than the lake's surface. Analyses of many samples showed that the ground water had an almost uniform isotopic composition, with a mean δ^{18}O value of -6.0‰. This proved that the lake, which had a mean δ^{18}O of $+0.6$‰, was not contributing significantly to the aquifer (Gonfiantini *et al.*, 1962).

The springs in the region of Lake Chala (Kenya) were also found not to be connected with the lake (Payne, 1970).

In the region of Lake Chad, Africa, samples of ground water collected at different depths, near to the shore line, showed important contributions of heavy-isotope-enriched lake water with more or less regular variations with depth and with distance from the lake. In principle, this makes it possible to evaluate in each point the proportions of lake water and of water from direct infiltration of precipitation (Fontes *et al.*, 1970).

In the region of Antalya (Turkey) some large karstic springs near the coast were believed to be fed by inland lakes, in the plateau at the north of the Taurus Mountains, which were known to leak important amounts of water through sinkholes and fractures. The fraction of water lost by such leakage was evaluated on the basis of an isotopic study of the lake water (Dinçer, 1968). The ^{18}O and D analyses of the spring waters showed not only that the contribution from the lakes was almost negligible but also that the recharge area of the springs was located on the southern side of the Taurus Mountains. In fact, crossing the mountains there is a change in the value of the intercept of the meteoric water line on the δD-axis (Dinçer and Payne, 1971).

10.2.5.2 Interconnexions between ground water and rivers

In the region of Gorizia (Italy), the two main rivers, Isonzo and Vipacco, descending from the eastern Alps, have waters with a δ^{18}O value of -10.5‰. The wells in the unconfined aquifer in the same region have values of δ^{18}O from -7 to -10.4‰, the values being more negative in proximity of the rivers and becoming more positive at increasing distances. It was thus possible to define the areas of infiltration of river waters in the unconfined aquifer and to determine its fraction with respect to local precipitation (Morgante *et al.*, 1966).

The influence also of the Chari river (Chad) (IAEA, 1972*a*) and the Nile river (Sudan) on the surrounding ground water has been shown by studies made at the Section of Isotope Hydrology of IAEA.

10.2.5.3 Interconnexions between aquifers

The large aquifer known as the 'Continental Intercalaire', (Lower Cretaceous) which is found in all Western Sahara (Africa) contains water with a uniform stable isotope composition throughout, and practically no ^{14}C. In the region contiguous to the unconfined aquifer of Grand Erg Occidental, the stable isotope composition of Continental Intercalaire water changes, and its ^{14}C content increases. This demonstrates clearly an important contribution from the Grand Erg Occidental aquifer, whose water is relatively recent and has a piezometric surface higher than that of the Continental Intercalaire (Conrad and Fontes, 1970, 1972; IAEA, 1972*b*).

In Tunisia, the Continental Intercalaire discharges in the overlying confined aquifer (coastal aquifer of Djeffara) through a fault system in the region of El Hamma west of Gabès. This is beautifully confirmed by the stable isotopes: the water in the coastal aquifer at El Hamma has exactly the same isotopic composition as that from the Continental Intercalaire, confirming its origin from it. The proportion of water coming from the Continental Intercalaire gradually decreases in the coastal aquifer from El Hamma to Gabès, in the direction of water flow (IAEA, 1972*b*).

A project in the Hodna region of Algeria concerns a study of the ground water in two aquifers in the region of a dry salt lake, Chott-el-Hodna, where flood waters from the near-by mountains evaporate completely. The heavy isotope content of water in the deeper aquifer is lower than that in the shallow aquifer, despite the fact that the two aquifers are presumed to have been recharged from the same general area. However, carbon-14 measurements giving ages of the order of some 30,000 years for the deep aquifer, indicate that the difference in stable isotopic composition is most probably due to an age effect (climatic change since the time of recharge). The deep aquifer is connected with the shallow one through buried alluvial fan (revealed by geophysics). In this area, the ^{14}C content of water in the deep aquifer increases together with the ^{18}O and D content. This is a clear indication of slow recharge from the shallow aquifer. In this study, tritium measurements also demonstrated the occurrence of recharge of the shallow aquifer from wadis (IAEA, 1972*c*).

10.2.5.4 Transit time and origin of water in aquifers

Stable isotope and tritium analyses were used in the volcanic island of Cheju (Republic of Korea), to characterize ground waters with respect to time and place of recharge and to determine the nature of mixing of the different ground-water sources and estimate their residence times. From a preliminary sampling of fourteen springs, eight wells and two streams from different parts of the island, nine points were chosen for periodic sampling. All the waters sampled contained appreciable amounts of thermonuclear tritium thus indicating rapid circulation in the various ground-water systems. Upon the basis of their tritium content and stable isotopic composition, the waters were classified into a number of flow régimes.

Springs at medium and high altitude exhibited a relatively high tritium concentration and a wide spread of stable isotopic composition. This was interpreted as the water having a short transit time with poor mixing, in general agreement with the hydrogeology of the area. On the other hand, large coastal springs had a much lower tritium concentration, but with a similar spread in stable isotopic composition. This suggested a similar source as for the high altitude springs but involving a somewhat longer transit time. Small springs and wells near the east coast of the island had a similar tritium concentration but, at the same time, somewhat higher heavy isotope content. The latter appeared to reflect the somewhat lower terrain in the eastern part of the island.

Although the period covered by the tritium analyses was relatively short, an estimate was made of the mean transit time of the waters. A well-mixed reservoir model was adopted assuming that recharge was only effective for monthly precipitation values in excess of 100 mm. Values of from 2 to 8½ years were estimated, the longer times being associated with waters discharged from an extensive well-mixed fresh-water lens (Davis *et al.*, 1970).

The recharge area and the turnover time (8 ± 3 years) has been determined for the Évian aquifer (Haute Savoie, France) with the help of stable isotopes and tritium. The results are in good agreement with the hydrogeological data (Fontes *et al.*, 1967).

A project in the coastal plain of Nicaragua concerns the study of recharge mechanism to a phreatic aquifer. The project area is relatively small, between the Pacific coast and the drainage divide of the volcanic chain, La Cordillera de Marrabios. Across the short dimension, elevations rise uniformly to about 200 m above sea level at about 20-25 km from the ocean, and abruptly thereafter to the Cordilleran crest which has an average elevation of about 1,000 m and a maximum of 1,745 m. The plain is covered by Quaternary *tobas*, ash deposits and the stratigraphic sequence of the Tertiary. Stable isotopes and tritium have been employed to study the problems associated with the recharge mechanism to the ground-water system in the area with particular emphasis on the importance of low plain recharge versus the elevated slopes of the Cordillera. Periodic sampling at five precipitation stations and two springs, all located at different elevations, enabled the establishment of the relation between the elevation and stable isotope composition of recharging water. Both shallow and deep wells in the plain are sampled periodically to determine the isotopic composition of local recharge as well as that of the deeper ground water. The rather shallow wells on the plain with most enriched heavy isotope content and highest tritium concentration were good indices for the isotopic composition of local recharge. The rather uniform stable isotopic composition of drilled wells, together with the known isotopic composition of local recharge and of the water recharging from elevated slopes, enabled the study of the relative proportion of plain recharge versus the recharge from elevated slopes. It was found that a much higher proportion of recharge to the ground-water system occurs from the elevated slopes with a mean altitude of about 500 m. The tritium content of wells in the plain enabled the study of the transit time involved in the recharge to ground water and it was concluded that the water in the shallow wells had been recharged locally no more than three years ago and the water tapped by the drilled wells had a mean transit time in the order of 20 years (IAEA, 1971*b*).

10.2.5.5 Flow velocity from ^{14}C data

An example of the field application of carbon-14 is provided by a study in Texas of a large confined aquifer. Estimated ages adjusted by the $\delta^{13}C$ approach ranged from zero in the recharge area to about 30,000 years. The estimated flow velocity of 1.5 to 2 m per year was in good agreement with that obtained from hydrological data (Pearson and White, 1967). A similar study was undertaken in the Ocala limestone aquifer in Florida where the estimates of ground-water velocity again agreed with hydrological estimates (Hanshaw *et al.*, 1965).

The use of carbon-14 for estimating the velocity of ground-water flow in an artesian aquifer near the coast in South Africa has also been reported. A fairly uniform increase in age from the outcrop area was indicative of relatively uniform rate of flow over the whole area of about 0.7 m/y (Vogel, 1970).

Such demonstration studies have given confidence in the validity of the ^{14}C method for estimating the flow velocity of ground water under favourable conditions. A major advantage of the ^{14}C method is that a mean regional velocity is obtained in contrast to the point values obtained in pumping tests.

10.3 Artificial isotope hydrology

Artificial radioisotopes can be measured in extremely low concentrations and often *in situ*, making possible the design of convenient and efficient field experiments. However, radioisotopes present the problem of health hazard, which increases cost due to the extra care needed in their use and handling. In some cases, the use of a radioisotope may present an unacceptable health hazard. Even where it can be demonstrated that the hazard to health is non-existent, local residents may oppose its use. Therefore, before considering the use of an artificial radioisotope as a tracer, in present or potential water supplies, the hydrologist should assure himself that other, non-radioactive substances, such as dyes or chlorides, would not meet the needs of his problem.

Radioactive isotopes are used as tracers to determine local characteristics of aquifers. The following parameters of an aquifer can be measured:

1. Aquifer characteristics (a) porosity, (b) transmissivity, (c) dispersivity.
2. Direction and velocity of ground-water flow.
3. Stratification of aquifers.

10.3.1 *Radioactive tracers*

The choice of the artificial isotope to be used depends on the features of the problem. In general, the following points should be taken into consideration:

1. The isotope should have a life comparable to the presumed duration of the observations. Unnecessarily long-lived isotopes will pollute the water, creating a persistent health hazard and interfering with repetition of the experiment.
2. The isotope should not be adsorbed by the mineral or organic aquifer components.
3. For most problems, it should be possible to measure the activity in the field. For this reason γ emitters are, in general, used.
4. The isotope must be available when and where required at a reasonable cost.

Some of the most frequently used radioactive tracers are listed in Table 10.3.1.

The chemical form of the radioactive solution used in water tracing experiments plays an important role. In many cases, cations serve as tracers. In the form in which they are commonly delivered, they may be subject to strong adsorption in the ground, interfering with the validity of their tracing of the water mass. To minimize this effect within an aquifer, tracers in a complex form are often used. The most commonly used in hydrological practice is a chelated metallic compound formed by means of ethylene-diamine tetracetic acid (EDTA).

Tritium, although not meeting the half-life and radiation requirements mentioned above, is an ideal marker for water since it is incorporated in the water molecule. Hence it is used in spite of the fact that *in situ* detection is not possible and samples must be analysed in the laboratory subsequent to an experiment. In certain cases HTO may suffer some delay in comparison with water bulk velocity as a result of ion exchange with clays (Kaufman and Orlob, 1956; Knutsson and Forsberg, 1967). The use of artificial tritium should be kept on the lowest possible level to minimize interference with ground-water resource studies based on environmental tritium.

10.3.2 *Techniques*

The introduction of tracer into a borehole may be done by pouring through a thin pipe, by crushing an ampoule at the depth of interest, or by using a special injection device.

TABLE 10.3.1. Radioisotope tracers used in ground-water investigations (Zuber, 1971; IAEA, 1966)

Isotope	Chemical form	Half-life	Decay modes	γ-energies in MeV and percentage of γ-quanta per disintegration	Maximum permissible concentration in drinking water (μCi/ml)	Minimum amount detectable in water (μCi/ml)
^{3}H (T)	H_2O	12.26 y	β^-	No γ	3×10^{-3} (10^{6} T.U.)	10^{-6} (300 T.U.)*
^{24}Na	Na_2CO_3	15.0 h	β^-	1.37 (100%), 2.75 (100%)	2×10^{-4}	10^{-8}
^{51}Cr	Cr-EDTA, $CrCl_3$	27.8 d	EC	0.324 (9%)	2×10^{-3}	8×10^{-7}
^{58}Co	Co-EDTA $K_3[Co(CN)_6]$	71 d	EC (85%) β^+ (15%)	0.81 (100%), 0.511 from positron annihilation	10^{-4}	6×10^{-8}
^{82}Br	NH_4Br	35.7 h	β^-	0.55 (69%), 0.61 (43%), 0.69 (29%) 0.76 (86%), 0.82 (26%), 1.03 (31%) 1.30 (31%), 1.46 (16%)	3×10^{-4}	2×10^{-8}
^{110m}Ag	$K[Ag(CN)_2]$	249 d	β^- (98%) IT (2%)	0.66 (100%), 0.88 (80%)	3×10^{-5}	3×10^{-8}
^{131}I	KI	8.05 d	β^-	0.08 (2.2%), 0.28 (5.3%), 0.36 (82%), 0.64 (9%), 0.72 (3%)	2×10^{-6}	8×10^{-8}
^{198}Au	$AuCl_3$	64.8 h	β^-	0.41 (99%), 0.68 (1%)	5×10^{-5}	10^{-7}

* Detection level without isotopic enrichment.

Injection can be performed at one or several depths to facilitate mixing of the tracer solution with the standing water of the borehole. After the release and mixing, the radioactivity in the borehole is measured by a probe, usually a scintillation counter but sometimes a rugged Geiger counter, inserted at the chosen depth. Specialized probes have been constructed which seal off a defined volume of the borehole by inflatable bladders, release radiotracer into this volume and measure the radioactivity by detectors in or immediately above the release volume (Fig. 10.3.2).

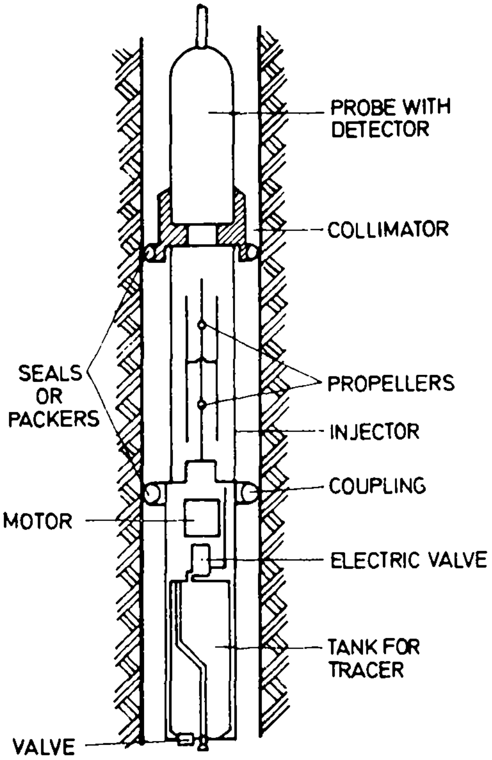

FIG. 10.3.2. Instrument for borehole dilution measurements (Guizerix *et al.*, 1963).

In a technique used by Kaufman and Todd (1962) tagged water was circulated between an isolated segment of a borehole and the ground surface by a small pump. Thus, continuous gentle mixing was obtained and injection and measurement of radioactivity were done at the surface without the necessity for a special probe.

In the Single Well Dilution Technique (Halevy *et al.*, 1967), the dilution rate of the tracer by the natural flow of water through the borehole is observed. In the Single-well Pulse Technique (Borowczyk *et al.*, 1967), tracer is forced into the aquifer by pumping water into the well and then recalled by pumping water out of the well.

When movement of water between boreholes is to be observed—Multiple-well Technique—the observation well is usually pumped and the radioactivity of the pumped water monitored at the surface (Halevy and Nir, 1962). It is important that such pumped water be disposed of in such a way that it cannot return inadvertently to the system under investigation. In some experiments, water pumped from the observation well is returned to the injection well intentionally to establish a closed circuit.

In shallow unconsolidated formations, measurement points may be established by driving a pipe into the ground with a vibration hammer. The maximum depth of penetration depends on the character of the materials and varies from 5 to 20 m.

10.3.3 *Applications*

10.3.3.1 **Effective porosity of an aquifer**

The principle of the method of porosity determination in the saturated zone is based on the approximate equality of porosity p (Void volume/Total volume) and partial volume of water s = Volume of water/Total volume.

Tracer is introduced into a well and a second well, at a distance r, is pumped. Disregarding the dispersion of tracer on its path between wells, its arrival at the second well signifies that the volume of water pumped, V, is

$$V = \pi r^2 bs \qquad 10.3(1)$$

where b = aquifer thickness, r = distance between wells, and s = effective porosity. Prime requirements are that the distance between wells be large compared to the aquifer thickness, ($r > b$), radial pumping velocities be larger than the natural velocities, and the cone of depression at the pumping well be small compared with the volume of water pumped.

A typical experiment of porosity determination has been fully described (Halevy and Nir, 1962).

10.3.3.2 **Transmissivity**

The coefficient of transmissivity T (or transmissibility), first introduced by Theis in his non-equilibrium equation, characterizes the ability of the aquifer to transmit water (Hantush, 1963). Along with the storage coefficient, hydraulic conductivity, specific storage and specific yield, transmissivity is one of the formation parameters characterizing basic hydraulic properties of the aquifer. It is related to the hydraulic conductivity K (mean filtration coefficient) and aquifer thickness b as

$$T = bK. \qquad 10.3(2)$$

Mercado and Halevy (1966) have found that the volume of water pumped during the experiment referred to in 10.3.3.1, measured until the tracer peak arrives at the next well, is inversely proportional to the value of transmissivity. They have determined transmissivity of two layers of the same aquifer by means of two injection and one observation well. Transmissivity then was found from the relationships:

$$\left.\begin{aligned} V_1 &= \pi r_1^2 b_1 s_1 T/T_1 \\ V_2 &= \pi r_2^2 b_2 s_2 T/T_2 \\ T_1 + T_2 &= T \end{aligned}\right\} \qquad 10.3(3)$$

where T = total transmissivity; $T_{1,2}$ = partial transmissivity of the layers 1 and 2; $r_{1,2}$ = distances between injection and observation wells; $b_{1,2}$ = thickness of the layers 1 and 2; $s_{1,2}$ = effective porosity of the layers 1 and 2.

Precision obtained was 5.10^{-6} m^2/s.

10.3.3.3 **Dispersivity**

In problems of artificial recharge of aquifers by water of a lower quality, or putting wastes into the ground water, a subject of interest may be the coefficient of dispersivity,

D, occurring in the tracer transport equation (Halevy *et al.*, 1967). This value characterizes the mixing property of an aquifer. It is commonly estimated by finding the theoretical curve which best fits the experimental tracer breakthrough curve in an observation well. Parameters on such theoretical curves are dispersion coefficients derived from the mathematical model of the tracer transport (Lenda and Zuber, 1970).

10.3.3.4 Ground-water flow velocity

Ground-water flow velocity under the natural hydraulic gradient is generally evaluated from Darcy's formula. Using a tracer, it is possible to measure directly the ground-water velocity, and from this to evaluate the aquifer permeability. The method consists of injecting tracer into the well and then following its concentration decrease with time (Fig. 10.3.3.4a).

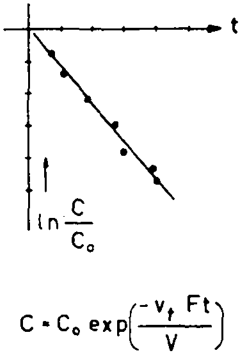

FIG. 10.3.3.4a. Schematic diagram showing how the tracer concentration changes with time in a borehole dilution experiment (Moser *et al.*, 1963).

The principle applies for any tracer, but again radioactive isotopes are easier to detect *in situ* with a high degree of accuracy, at very low concentrations (Moser *et al.*, 1957, 1963; Mairhofer, 1963; Guizerix *et al.*, 1963).

Disadvantages of the method are that the results are valid only in a restricted area neighbouring the well. Several measurements in different wells and at different depths are often necessary to establish a good picture of the ground-water flow in a given aquifer.

The horizontal filtration velocity V_f of water (sometimes also called Darcy velocity) is given by:

$$V_f = -\frac{V}{\alpha F f}\ln\left(\frac{C}{C_0}\right) \qquad 10.3(4)$$

where V is the measuring volume (the borehole volume in which dilution takes place), F is the cross-section of the measuring volume perpendicular to the direction of the undisturbed ground-water flow, and t is the time interval between measurement of concentrations C_0 and C; α is a correction factor accounting for the distortion of the flow lines due to the presence of the borehole. Calculation of the value of α on the basis of well-construction data is discussed by Halevy *et al.* (1967). The above expression is a particular solution of the differential equation describing dilution rate of the tracer.

In practice, several readings are required to derive results not affected by incomplete mixing processes and residual currents. The velocity lower limit, determined mainly by the diffusion rate, is 10^{-7} m/s.

If effective porosity, s, is known one can estimate the bulk velocity as:

$$V_{\text{bulk}} = \frac{V_f}{s}. \qquad 10.3(5)$$

Potential (hydraulic) gradients, dh/dx, can be determined from the vertical distribution of radioactivity in the observation holes if these boreholes are entirely cased (Weinberger *et al.*, 1967). Knowing the potential gradient, one can also deduce the so-called field filtration coefficient (or hydraulic conductivity, or permeability) K, applying Darcy's law:

$$K = \frac{V_f}{\frac{dh}{dx}} \qquad 10.3(6)$$

Knowledge of the vertical velocity component in a borehole is important in such problems as determination of the amount of water entering or leaving the well from different zone locations of permeable strata, or amount of interchange between zones. A single well-dilution technique may be applied. A series of detectors placed at different heights is commonly used for such measurements (Fig. 10.3.3.4b). The method may be specially recommended when expected vertical flow velocities are smaller than 2.10^{-2}

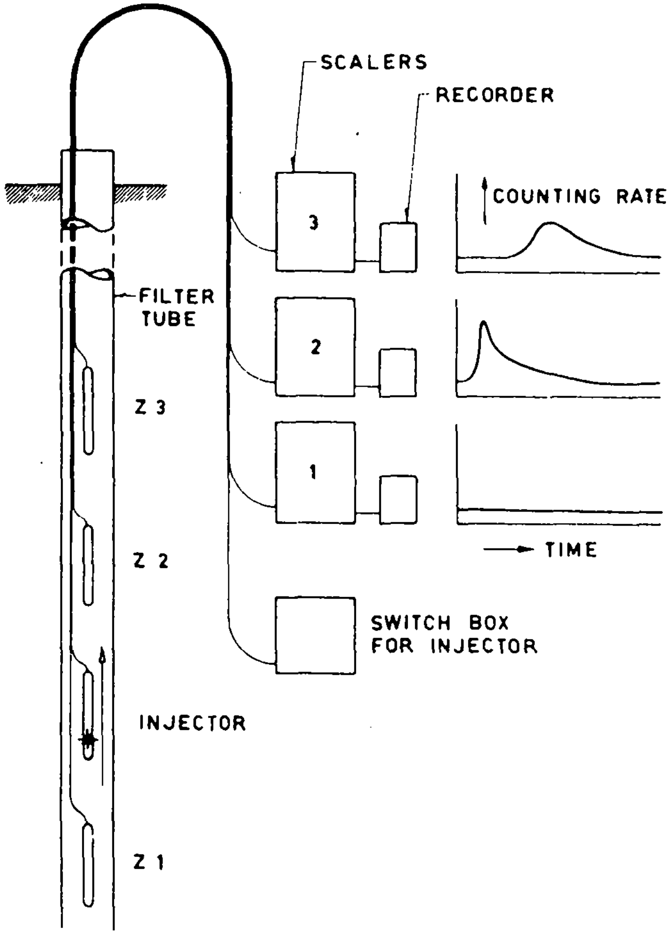

Fig. 10.3.3.4b. Arrangement of a set of detectors for determining vertical flow in a borehole (Moser *et al.*, 1963).

m/s; in this region, due to the influence of friction, tracer methods are more efficient than the use of mechanical current meters. The lower velocity limit is estimated at 10^{-7} m/s. In turbulent flow, accuracy of measurement is reported to be better than 10 per cent but less is known about the accuracy in laminar flow (Halevy *et al.*, 1967).

10.3.3.5 **Direction of ground-water flow**

Multiwell technique can be used to determine the direction of ground-water flow, especially when radial symmetry in well location exists, or if observation wells are found in the approximate direction of flow. Direction of flow, however, may also be determined by the means of a single-well technique. Tracer is added to a segment of a borehole and is carried away in the direction of the flow. Activity is then detected by a special directionally oriented probe rotated by means of a stiff metallic rod. A diagram like that of Fig. 10.3.3.5 is then obtained, which shows the direction of the water flow. In this case, contrary to the flow-velocity measurements, radioisotopes that are relatively rapidly absorbed by the formation are the most appropriate ones to use. Gold-198 chloride ($^{198}AuCl_3$) and chromium-51 chloride ($^{51}CrCl_3$) are commonly used. In most laboratory tests the difference between measured and the true flow direction was less than 3°. The reproducibility of the method in the field is better than ± 10 per cent. A simplified version of the method has been used by Wurzel and Ward (1965). A cylindrical metal gauze placed in the well at the injection point adsorbs radioactive tracer ($^{51}CrCl_3$). The gauze is removed, cut in many sections parallel to the axis and the activity of each section is measured in the laboratory. The section giving the highest activity indicates the direction of water flow.

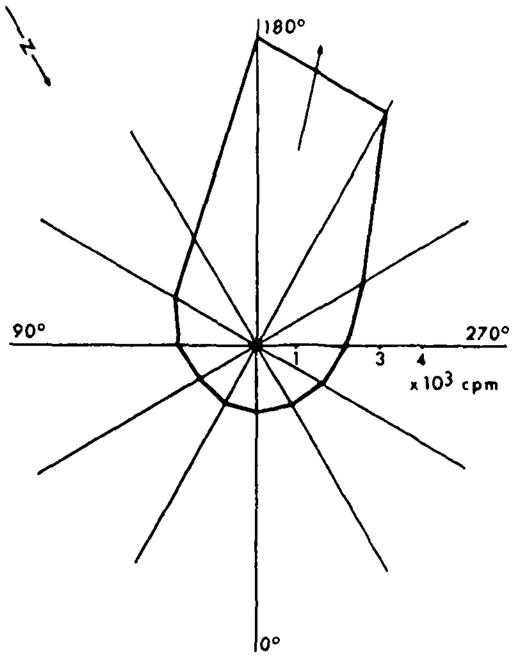

FIG. 10.3.3.5. Diagram of water flow direction determination in a borehole. Activity injected, 0.6 mCi (Halevy *et al.*, 1967).

In a similar method proposed by Hazzaa (1970), the radioactive tracer ^{32}P (half-life 14.3 d) is adsorbed on two coaxial metallic screens. Subsequent beta radiographic examination of both screens yielded information on the basis of which filtration velocity and direction of flow could be determined. Until now only results of the laboratory tests are available.

10.3.4 *Practical considerations*

Some authors are very enthusiastic about the possibility of radiometric methods (Feely *et al.*, 1961), whereas others emphasize that conditions existing in real aquifers make results of the tracer measurements doubtful. Sometimes, complexity of the aquifer does make a reasonable interpretation of the tracer data impossible, but such cases fortunately are rare. In some cases radiometric methods offer sensitivity and accuracy of measurement higher than the classical ones. Sometimes, they are the only ones available in solving specific problems.

Since commercial service in the application of these techniques is seldom available, it is advisable to seek the co-operation of universities or research institutes in the initial use of the methods. Some difficulties in performing field measurements may be expected until the experimenter becomes experienced in the new techniques involved. Some of the radiometric methods are still in the development stage and simplification and improvements in technique can be expected in the future.

A typical list of experimental details to be considered specifically is as follows:

1. Parameters to be determined.
2. Selection of tracer, amount required, means of producing it at the required time.
3. Provision of requisite injection and detection equipment and pumping equipment of adequate capacity.
4. Means of disposal of pumped water to prevent its recirculation to the borehole.
5. Transportation to and temporary storage of the radioactive tracer at the experimental site.
6. Radiological safety of the operating personnel.
7. Ensuring that no significant amount of tracer will get into public water supplies.
8. Evaluation of the results.

The experimenter must acquaint himself with all safety and legal requirements connected with the handling of radioactive materials. The International Atomic Energy Agency's *Guide to the Safe Handling of Radioisotopes in Hydrology* (1966), serves as a very useful introduction to this subject.

10.4 Glossary

^{14}C age, adjusted. The age calculated from the ^{14}C content of the sample and from an initial ^{14}C content evaluated on the basis of a given correction method: for most of the authors, on the basis of the $\delta^{13}C$ value of dissolved carbonate species.

^{14}C age, apparent. The age calculated from the ^{14}C content of the sample, assuming an initial ^{14}C content equal to 100 per cent of the modern (condition which is never realized for carbonate species dissolved in water):

$$\text{Apparent age} = 8{,}270 \ln \frac{100}{C} \text{ years}$$

where C is the ^{14}C content of the sample relative to modern.

^{14}C Concentration Unit. The ^{14}C content of natural carbon compounds is expressed in per cent of the ^{14}C content of 'modern' carbon. The latter has been conventionally accepted as equal to 95 per cent of the ^{14}C content (in the year 1950) of an oxalic acid standard distributed by the National Bureau of Standards, Washington, D.C., and corresponds to the normal ^{14}C content of plants before 1890.

Cosmic radiation. The primary cosmic radiation (originating in the sun, solar component, and in the stars, galactic component) consists of very-high-energy protons and α-particles, with a small fraction of heavier nuclei. Secondary cosmic radiation is produced by interaction of the primary cosmic radiation with the components of the upper atmosphere. It consists of a great variety of nuclear particles, among them neutrons, protons, mesons, hyperons, antiparticles, etc.

Tritium and ^{14}C are formed by reactions of the protons and neutrons with the atmospheric components. The rates of production are: for tritium, 0.25 atoms/cm^2/sec; for ^{14}C, 2.5 atoms/cm^2/sec.

Counter, Geiger-Müller. An instrument for the detection of radioactive substances and for determining their concentration. Basically, it consists of a gas-filled metallic cylinder, having a wire along its axis. The wire is about 1,000-1,500 V positive with respect to the cylinder walls; α, β or γ radiation entering the cylinder causes ionization of the gas and a discharge between the wire and the walls, which is detected and recorded with appropriate electronic instrumentation.

Counter, proportional. This counter is commonly used for tritium and ^{14}C determination. It is essentially the same as the Geiger-Müller counter, but the potential of the central wire is adjusted in such a way that the current pulses produced by the radiations are proportional to the energy released. Therefore, radiations having different energies can be distinguished and counted separately.

For tritium and ^{14}C determination in natural samples, the counter is filled with a gas (commonly ethane or methane) which has been synthesized with the hydrogen or the carbon to be analysed. The background counting rate of the counter is greatly reduced with a thick shield of lead or steel and with coincidence counters arranged to eliminate cosmic radiation from the recorded counts.

Counter, scintillation. This counter utilizes the fact that certain substances emit a light flash when they interact with a radiation. The light flashes are detected by a photomultiplier and recorded.

In isotope hydrology, scintillation counters are commonly used for detecting the γ-rays emitted by artificial isotopes. The scintillator is a large crystal of NaI, activated with traces of thallium iodide. A ratemeter gives the number of disintegrations per unit time, providing an immediate measurement of the radioisotope concentration.

Liquid scintillation counters are used for tritium and ^{14}C determination. In these counters a liquid that scintillates on interaction with a β-particle is mixed with the liquid under investigation (commonly water in the case of tritium, and benzene in the case of ^{14}C).

Curie (Ci). The unit for expressing radioactivity. It corresponds to the quantity of any radioactive isotope undergoing 3.7×10^{10} disintegrations per second.

Millicurie (mCi = 10^{-3} Ci), microcurie (µCi = 10^{-6} Ci), nanocurie (nCi = 10^{-9} Ci) and picocurie (pCi = 10^{-12} Ci) are normally used units.

Delta notation (δ‰). Defined by:

$$\delta‰ = \left(\frac{R_{\text{sample}}}{R_{\text{standard}}} - 1\right) \times 1000$$

where R = D/H, or $^{18}O/^{16}O$, or $^{13}C/^{12}C$. The standards generally adopted are: SMOW for hydrogen and oxygen isotopes and PDB (see below) for carbon. The corresponding notations are: δD‰, $\delta^{18}O$‰, $\delta^{13}C$‰. A δ‰ = + 10 (or − 10) means that the heavy isotope (D, or ^{18}O, or ^{13}C) content of the sample is higher (or lower, if the value is negative) by 10‰ than that of the standard. The usual accuracies in δ‰ determinations are: 1‰ for D/H and 0.1‰ for $^{18}O/^{16}O$ and $^{13}C/^{12}C$. Because of the fact that in natural waters the relative variations of the D/H ratio are generally five to eight times higher than those of $^{18}O/^{16}O$, the accuracies for both hydrogen and oxygen are in practice comparable.

Environmental isotopes. Those isotopes, of both natural and artificial origin, which occur in the natural environment on a global, or at least on a regional scale without any possible control by man. The variations of such isotopes in natural waters can be used by hydrologists for studying some important features of surface and ground waters such as origin of water, age and flow rate of water, etc.

The environmental isotopes most commonly used in isotope hydrology are: the stable isotopes of hydrogen and oxygen, the two constituents of water; ^{14}C and tritium, both of natural (cosmic rays) and artificial (thermonuclear tests) origin.

Environmental water tracer. Any substance appearing in natural waters, derived from the environment, whose concentration distribution in space and/or time can be used to differentiate or 'tag' specific water masses, or to identify the origin or provenance of specific water masses. Environmental tracers commonly enter the hydrologic cycle through atmospheric processes, but also may enter by leaching or exchange with rocks or organic material in the soil horizon, or with rocks at depth, or by radioactive decay of rocks at depth. Naturally occurring isotopes have certain unique properties that can be applied to the solution of hydrologic problems where ordinary tracers could not.

Evaporation line (stable isotopes). In a ($\delta^{18}O$, δD) diagram, a line of the equation:

$$\delta D‰ = a\, \delta^{18}O‰ + b$$

with $4 < a < 8$ and $b < 10$, fitting the points representing the isotopic compositions of waters which, having the same initial isotopic compositions, have undergone different degrees of free surface evaporation under similar environmental conditions.

The evaporation lines in a ($\delta^{18}O$, δD) diagram lie on the right side of the meteoric water line, the interception of the two lines giving the $\delta^{18}O$ and δD values of the water before evaporation.

Evaporated waters are most commonly found among surface waters and especially lake waters. Also ground waters, totally or partially originating from surface waters which have undergone intense evaporation, may show isotopic compositions which deviate from the meteoric water line and follow evaporation lines.

Half-life. The time necessary to reduce the concentration of a given radioactive isotope to half of its initial value by radioactive decay: $C = 1/2\, C_0$ (see below, 'Radioactive Decay Law'). The half-life is then:

$$t_{1/2} = \frac{\ln 2}{\lambda}$$

where λ is the rate constant characteristic of the decay of the isotope.

Isotopes. Atoms of the same chemical element with different masses are isotopes. Isotopes have the same number of protons in the nucleus but a different number of neutrons. For instance, oxygen has three natural isotopes with masses of 16, 17 and 18:

$^{16}_{8}O$: 8 protons + 8 neutrons in the nucleus

$^{17}_{8}O$: 8 protons + 9 neutrons in the nucleus

$^{18}_{8}O$: 8 protons + 10 neutrons in the nucleus

The lower index at the left of the chemical symbol indicates the number of protons in the nucleus (atomic number); the higher index indicates the sum of protons and neutrons (mass number). In the text, the atomic number is omitted, as isotopes are identified by the chemical symbol of the element and by their mass number.

Only isotopes of hydrogen have special names and symbols. H—protium, the most common isotope with mass 1; D—deuterium, the isotope of mass 2; T—tritium, the isotope of mass 3. Isotopes can be naturally occurring or artificially produced (e.g. in a reactor), stable or radioactive.

Isotopes, radioactive. Radioactive isotopes have unstable nuclei. They change into isotopes of other elements with time (radioactive decay). The rate of decay is generally expressed by means of the half-life or the mean life.

The concentration of radioactive isotopes in a closed system decreases exponentially with time unless there is production at the same time by some other radioactive processes.

Isotopes, stable. Stable isotopes have a nucleus with stable configuration. Their concentration in a closed system does not change with time unless they are produced by some radioactive element present in the system.

Isotopic exchange. The interchange of atoms between chemical compounds or phases. This process allows different isotopic compositions of co-existing compounds or phases to come into an equilibrium characteristic of the conditions of the system. For hydrological purposes the more important isotopic exchange processes are:

1. Exchange between water and rocks (especially limestone). In this process only the oxygen isotopic composition of water and of limestone can be affected (the hydrogen content of rocks is generally small). This process is very slow at the temperatures normally occurring in aquifers and it may be of importance only for thermal waters where the high temperature significantly increases the exchange rate.
2. Carbon isotope exchange between dissolved bicarbonate and solid $CaCO_3$ in the aquifer, which may modify the ^{13}C and ^{14}C contents of the bicarbonate used for ^{14}C age determination. This process, also, is very slow at normal temperatures, but the amount of dissolved bicarbonate is relatively small and its isotopic composition can be affected. In this case, the $\delta^{13}C$ of bicarbonate will be displaced towards more positive values.

Isotopic fractionation. Difference in the distribution of the isotopes of the same element in two chemical compounds or phases in mutual isotopic exchange. Isotopic fractionations occur because the isotopes have slightly different physical and chemical characteristics, due to their mass differences. Therefore, the isotopic fractionations are, in general, higher for light elements because the relative mass difference is higher.

Isotopic fractionation factor. The ratio between the isotopic ratios of two different compounds or phases in mutual isotopic exchange. The fractionation factors are in general indicated with α. For example, in the equilibrium water-vapour at 0° and 25° C the fractionation factors for oxygen and hydrogen isotopes are:

$$\alpha^{18}O = \frac{(^{18}O/^{16}O)\ \text{water}}{(^{18}O/^{16}O)\ \text{vapour}} \quad 1.0115\ (0^\circ\ C);\quad 1.0092\ (25^\circ\ C)$$

$$\alpha D = \frac{(D/H)\ \text{water}}{(D/H)\ \text{vapour}} \quad 1.108\ (0^\circ\ C);\quad 1.077\ (25^\circ\ C)$$

The fractionation factors tend to approach the value of 1 with increasing temperature. When $\alpha = 1$, the isotopic ratios are equal in both the compounds considered and, therefore, there is no isotopic fractionation. Sometimes fractionation is expressed as $\varepsilon‰ = (\alpha - 1) \times 1{,}000$.

Mass spectrometer. An instrument which allows the determination of the relative abundances of components having different masses in a given substance: as, for instance, in the case of isotopic species of a chemical compound.

In the most common type of mass spectrometer, the compound under investigation is ionized (e.g. by impact with an electron beam) and the ions are accelerated through an electric field. Then they enter a magnetic field normal to their path where they follow circular trajectories, the radii of which depend on the mass of the ions:

$$r = \frac{1}{H}\sqrt{\frac{2\,Vm}{e}}$$

(r = radius, H = magnetic field intensity, V = accelerating voltage, m = mass of the ion, e = electric charge of the ion).

Ions having different masses emerge from the magnetic field following different trajectories, so that they can be collected separately and their relative concentration can be measured.

In hydrology, mass spectrometers are used for measuring stable isotopes' abundances. The abundances of the radioactive isotopes of interest are too low to be measured with this instrument.

Mean life. The mean time τ of life of a given radioactive isotope:

$$\tau = \frac{\int_0^\infty C\,dt}{C_0} = \frac{1}{\lambda} = \frac{\text{half life}}{\ln 2}$$

(see below, 'Radioactive Decay Law').

Meteoric water line (stable isotopes). In a ($\delta^{18}O$, δD) diagram, the line of the equation:

$$\delta D\text{‰} = 8\delta^{18}O\text{‰} + 10$$

which has been found to best fit the points representing the isotopic compositions of precipitation samples from all over the world, at least for values of $\delta^{18}O$ lower than 0 (with respect to the SMOW).

In some regions of the world the straight line representing the correlation between $\delta^{18}O$ and δD is displaced, with generally the same slope 8 but a different value of the intercept on the δD axis (regional meteoric water line).

Ground waters and surface waters follow in most cases the same $\delta^{18}O$-δD relationship unless they have undergone substantial evaporation, as occurs often in lake waters. Free surface evaporation tends to enrich the heavy isotope content of water but not in the relative proportion established by the above relationship (see above, 'Evaporation Line').

PDB. $CaCO_3$ obtained from the calcareous rostrum of a Cretaceous Belemnite *(Belemmitella Americana)* from the Peedee Formation of South Carolina. Its carbon isotopic composition represents a good average of that of marine limestone. Its ^{13}C abundance is 1.111 per cent.

The PDB is used also as a reference standard for $^{18}O/^{16}O$ determinations in palaeotemperature measurements. The PDB is now no longer available. Other standards, whose isotopic composition with respect to PDB are known, are available for calibration: the NBS 20 (Solenhofen Limestone) distributed by the National Bureau of Standards in Washington, D.C., and the TS (Marble) distributed by IAEA.

Radioactive decay law. Radioactive decay follows the law:

$$\frac{dC}{dt} = -\lambda C$$

where C is the concentration of radioactive nuclei in the system at the time t, dC/dt is the decay rate and λ is the decay constant characteristic of each isotope. By integration we get:

$$C = C_0 e^{-\lambda t}$$

where C_0 is the concentration of radioactive nuclei present at the initial time ($t = 0$). Knowing C_0, λ, and C (the present concentration) we may evaluate the age of the system:

$$t = \frac{1}{\lambda} \ln \frac{C_0}{C}$$

Radioactivity. Radiation emitted when unstable isotopes are transformed (decay) into isotopes of other elements by nuclear processes. The most common processes of radioactive decay are:

1. *α-decay.* An α-particle, having mass 4 and charge 2 (a helium nucleus) is emitted by the nucleus of the parent isotope, which is thus transformed into an isotope of a different element having a mass number lower by 4 units and an atomic number lower by 2 units. Example:

$$^{238}_{94}U \xrightarrow{\alpha} {}^{234}_{90}Th$$

2. *β-decay.* An electron is emitted by the nucleus of the parent isotope, which is thus transformed into an isotope of a different element having the same mass number and an atomic number higher by 1 unit. Examples:

$$^{3}_{1}H \xrightarrow{\beta^-} {}^{3}_{2}He$$

$${}^{14}_{6}\mathrm{C} \xrightarrow{\beta^-} {}^{14}_{7}\mathrm{N}$$

3. *E.C. (electron capture) decay.* The nucleus of the parent isotope captures an orbital electron and is thus transformed into an isotope of a different element having the same mass number and an atomic number lower by 1 unit. Example:

$${}^{40}_{19}\mathrm{K} \xrightarrow{\text{E.C.}} {}^{40}_{18}\mathrm{Ar}$$

The product of the decay is usually in an excited state. It returns to the ground state by the emission of energy in the form of X- or γ-rays (electromagnetic radiations).

SMOW (Standard Mean Ocean Water). A standard water having an isotopic composition close to that of mean ocean water. The abundance of the three main isotopic species of water in SMOW are: $H_2{}^{16}O$ 99.73 per cent; $HD^{16}O$ 0.031 per cent; $H_2{}^{18}O$ 0.199 per cent. The advantages of using SMOW as reference standard for hydrogen and oxygen isotopic analyses of water are:

1. The ocean contains some 97 per cent of the water present on the earth's crust, and has a fairly uniform isotopic composition.
2. The ocean constitutes the initial and the final point of any important hydrological circuit. By definition, SMOW has δD and $\delta^{18}O$ equal to zero. The IAEA distributes samples of water having an isotopic composition practically identical to that of SMOW for purposes of calibration.

Tritium Unit (T.U.). The unit used for expressing the tritium concentration in natural samples. One T.U. corresponds to a concentration of one tritium atom per 10^{18} atoms of hydrogen. Conversion to μCi/ml:

$$1\ \text{T.U.} = 3.24 \times 10^{-9}\ \mu\text{Ci/ml}.$$

Further reading

Guidebook on nuclear techniques in hydrology, prepared by the Working Group on Nuclear Techniques in Hydrology of the International Hydrological Decade. Vienna, IAEA, 1968. 214 p. U.S.$5.50. (Technical reports series, no. 91.)

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Isotopes in hydrology, proceedings of a symposium held in Vienna in 1966. Vienna, IAEA, 1967. 740 p. U.S.$15.

Isotope hydrology 1970, proceedings of a symposium held in Vienna in 1970. Vienna, IAEA, 1970. 918 p. U.S.$24.

Isotope techniques in hydrology, vol. I (1957–65). Vienna, IAEA, 1968. 230 p. U.S.$6. (Bibliographical series, no. 32.) This book contains 985 references and summaries of papers in the field.

The use of nucleonic instruments for well logging is discussed in Chapter 3 of this guide. Definitions of specialized terms can be found in the Glossary, Section 10.4.

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14 Analytical and investigational techniques for fissured and fractured rocks[1]

14.1 The hydrogeology of fractured and fissured rocks

14.1.1 *Introduction*

The rocks discussed under this heading include indurated rocks, such as igneous, intrusive (plutonic) and extrusive (eruptive) types, metamorphic rocks and indurated sedimentary rocks (sandstones, quartzites, shales, etc.). The fractures and fissures in these rocks remain relatively unchanged as the water circulates, unlike limestones and dolomites in which changes do occur, the fissures and fractures growing larger owing to solution and erosion. These carbonate rocks will accordingly be dealt with separately (see 14.3).

The metamorphic and intrusive or plutonic rocks, which comprise continental shields as well as smaller or crystalline rock masses, lie at or near the land surface in about one-third of the area of the world's continents. Extrusive volcanic rocks also cover vast regions as, for example, in the Deccan of India, in Australia, in East Africa, in the United States, in South America and elsewhere. Carbonate and other indurated sedimentary rocks also figure importantly in sedimentary basins and mountain ranges. The need for study of their hydrogeology in order to understand the occurrence of ground water in a large part of the globe can thus be appreciated.

The study of the hydrogeology of fractured and fissured rocks differs greatly from the study of friable rocks with interstitial porosity. Hence it has been the subject of relatively little scientific work, and it occupies only a very small place in current textbooks. The fact is that many scientists lack a clear understanding of how to tackle hydrogeological problems in fissured and fractured rock terrains. The study of the hydrogeology of friable rock aquifers may begin with the measurement of porosity and permeability on a small volume of rock, the results of which can then be extrapolated to the whole aquifer. On the other hand, in the study of the hydrogeology of fractured and fissured rocks where the openings are widely separated, it may be necessary to change scale from a cubic decimetre of rock with interstitial porosity to a cubic hectometre or cubic kilometre of rock with fissure porosity. Broad-scale field-work thus may take the place of detailed laboratory work, and geological study has priority. Just as the study of pores and the water circulation within them is essential for the understanding of water circulation in friable rocks, so the study of water circulation in fissures is essential for an understanding of water circulation within a mass of indurated rock.

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[1] This chapter has been prepared by Dr H. Schoeller on behalf of the *ad hoc* subgroup of the IHD Working Group on Ground-water Studies and revised by a panel of editors composed of: A. A. Konoplyantsev, E. P. O'Driscoll, H. J. Schoeller and G. C. Taylor.

14.1.2 *Fractures and fissures*

Some fractures and fissures have general characteristics common to all categories of rocks, but there are some with characteristics peculiar to sedimentary rocks, to metamorphic rocks, to granites or to volcanic rocks. Only fractures and fissures common to all categories of rocks are dealt with here. Separate sections describe the hydrogeology of the fractures and fissures of each of these rock categories.

The terms 'fracture' and 'fissure' are often treated as opposites in textbooks on geology. In this text, the discussion is limited to the meaning given in technical dictionaries, that is, a fracture is the result of a violent break requiring energy, and a fissure is a small tabular parting, in other words, a narrow opening of varying length, irrespective of its origin.

Faults are long, deep fractures or fracture zones, with a vertical or oblique attitude, separating two rock units which have undergone relative displacement. They are commonly some distance apart and parallel to each other, thus forming wide or long, vertical or sub-vertical units. They are at places intersected by networks of other fractures forming vertical or sub-vertical blocks. In neither case are these units and blocks common to another fault system either horizontally or obliquely. Even faults which run in the same direction may intersect each other at depth by reason of a difference in inclination to the vertical. Thus, faults may act as drains for intersected aquifers; as sub-surface dams to the aquifers by juxtaposing impermeable barriers to the flow of water through aquifers; and as siphons. Water circulation through faults or fault zones is not uniform and is frequently localized, because faults are not necessarily open at all places or in all directions.

A normal fault is a fracture caused by tension or stress as a result of which two adjoining rock masses have generally moved apart vertically and horizontally and have undergone some relative displacement, commonly with accompanying crushing of the rocks adjacent to the fault. Another thrust is a type of fault separating two rock masses, one of which overrides the other for some distance. As it is the result of compressive stress the fault zone tends to be tight because the two sides are pressed together, especially in the case of a high-angle fault. In faults, however, the crushed zones may become channels for ground-water movement.

Joints are partings in a rock mass which generally result in an increase in rock volume without any relative movement parallel to the places of the joints.

Tectonic joints result from strain placed on rocks during orogenic movements. A distinction is made between shear joints which are closed; tension joints which tend to remain open and can thus play some part in allowing ground-water movement when they are intersected by other joints, fissures or open fractures; exfoliation joints which occur near the land surface and result from weathering processes, including changes in temperature, freezing and thawing, hydration or crystallization of rock minerals and the growth of roots, and which are widened by physical, chemical and biological degradation. They most frequently originate from one of the various types of fissures.

Pressure relief fractures or sheet joints form in the rocks along the sides and underneath valleys as a result of removal of rock load which occurs when the valleys are formed. These fractures mostly run subparallel to the sides and floors of the valley and extend as much as 100 and 200 m upslope in the valley walls, depending on the kind of rock and the degree of slope. Such joints may be from 1 mm to as much as 5 cm wide.

The degree of fracturing depends on the intensity of the tectonic stresses which created the fractures. Some zones have undergone almost no movement, while others have been subject to considerable compression or decompression, particularly adjacent to mountain ranges.

The degree of fracturing may be still greater in a region which has been affected by several orogenies producing an intersection of fractures which is favourable to the circulation of water.

The size and the extent of fractures enlarged by weathering depends, of course, on the climatic conditions and the length of time the rocks have been exposed to these conditions. To give an example, a crystalline rock mass planed by ice action displays a low degree of open fracturing, whereas a crystalline rock mass which has been exposed to long-continued continental weathering may have a thick cover of disintegrated rock.

The depth to which fracturing extends depends on the origin of the fractures. Faults may be deep and some faults may even pass through the entire thickness of the earth's crust. Joints, on the other hand, may only permit movement of water to depths of about 100 m or less. This situation will be dealt with for each category of rock. The depth of fractures enlarged by weathering may be anything from a centimetre to tens of metres.

14.1.3 *Porosity*

The interstitial porosity of compact rocks is generally low, ranging from almost nothing to a few per cent. Limestones may have porosity equal to that of friable rocks, in certain cases.

Generally, fissure porosity is important in compact rocks and can be measured only for a large volume of rock. The difficulty of the measurement is obvious. It would be possible to count the number of fissures in a given volume of rock, but more difficult to determine their dimensions.

14.1.4 *Hydrodynamics of fissures*

Depending on the width of the fissure and the rate of movement of water, flow may be either laminar or turbulent, or in a transition zone with a high Reynolds number.

In laminar flow, the discharge per unit of fracture length measured perpendicular to the flow, is as follows:

$$q = \frac{l^3 \gamma I}{12\delta} = \frac{l^3}{12\nu} I \text{ (cm}^3\text{/cm}^2\text{/s)}$$

where:

I = the hydraulic gradient along the plane of the fracture;
l = the width of the fracture;
γ = the specific weight of the water;
δ = the dynamic viscosity of the water;
ν = the kinematic viscosity of the water.

This leads to: $q = KlI$, with $K = \frac{gl^2}{12\nu}$.

In a turbulent flow: $q = K'l\sqrt{I}$. Volod'ko laboratory experiments (Volod'ko, 1941) are shown in Table 14.1.4a.

TABLE 14.1.4a. Relation between water flow and width of fractures

Rock	Width of fractures (cm)	Maximum velocity without turbulence (cm/s)	Rock	Width of fractures (cm)	Maximum velocity without turbulence (cm/s)
Slaty schist	0.001–0.01	No turbulence	Granite	0.5	10.0
Sandstone	0.3	1.0	Basalt	0.5	0.1

In practice, there is no flow which conforms wholly to either the linear or the quadratic relationship, because in narrow joints the flow is laminar, while in the wide ones it has high Reynolds numbers; near the point of entry into the opening the velocity may be high, decreasing with distance as the cross-sectional area becomes large.

It is advisable therefore to look for a formula which is valid without any restriction of velocity or Reynolds number, such as:

$$I^{3/4} = \left(\frac{v}{K}\right)^{3/4} + \left(\frac{C}{100\,g\,l}\,v^2\right)^{3/4}$$

where:

I = the hydraulic gradient;
v = the mean velocity of the fissures;
K = the hydraulic conductivity;[1]
g = the acceleration due to gravity;
l = the width of the fissure;
C = a constant between 6 and 30 depending on the roughness of the walls and the density of the fissured network.

The volume of water which can be conveyed by one single fissure can be used in calculating the parameter of the hydraulic flow through a fissured layer. They can be statistically related. In this it is an essential prerequisite that the opening sizes should have random values in the rock area being investigated. However, if this area is small then the surface openings, by which is meant the proportion of the area of openings to the total area (F_a/F), may change in magnitude depending on the location of the selected unit area. If this surface area is large enough, the proportion of surface openings has a well-defined mean and an acceptable small variance.

If the opening sizes have a random value, the widths of the fissures should be plotted as a statistical distribution curve, which will show the probability of occurrence of a fissure of any given width. Knowing that the volume of water conveyed by a fissure is a function of its width, the statistical distribution of the flow rates through the different openings can also be described. In this way the total flow through a unit area can also be calculated.

14.1.5 *Permeability*

The permeability of the fracture system of the rock formation as a whole is of more interest than the ability of an individual fracture to transmit water. The permeability cannot be estimated, of course, unless the mass of rock is sufficiently large. The permeability of the mass as a whole depends on the collective permeability of each of the fractures of an interconnecting system, in other words on the extent to which they are open, on the number of fractures per unit of volume and, especially, on the degree of intercommunication between fractures.

Only rough estimates and comparisons are usually possible. One rock mass may be more permeable than another because the fractures are more open and more numerous as a result of its being more affected by tectonic movements. An opinion may be based only on field observations. When fractures are numerous and well interconnected, pumping tests can provide approximate values of permeability.

14.1.6 *Ground-water bodies*

The pattern of fracturing controls the accumulation of ground water. If the fractures are unidirectional, an interconnected ground-water body will not form, and isolated

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[1] The terms 'permeability' and 'hydraulic conductivity' are used synonymously in this chapter.

water-filled cracks and resources will provide only a limited water resource. If the fractures intersect, and particularly if the fracture network is dense, there may be good intercommunication. All variations are possible between these two extremes.

14.1.7 *Recharge*

The weathered zone is highly important because where it is permeable, it may accumulate and store water infiltrated from the surface. In the absence of weathering, the water runs over exposed rock surfaces and only a small amount can penetrate into the fissures.

The ability of fissures to accept water infiltrating depends essentially on the absorption capacity (C_a) of the fissures which in turn depends on the transmissivity (T) and on the hydraulic gradient (I) between the zone of infiltration and the discharge zone ($C_a = TI$) (see 14.3.3.4). Replenishment thus depends on the geometry of the fissures.

14.2 The hydrogeology of non-carbonate hard rocks

14.2.1 *Intrusive igneous rocks*

14.2.1.1 **Introduction**

This section includes granite, syenite and diorite. The hydrogeology of granite massifs is best known.

Although granites are commonly massive, joints may develop in three ways: (a) joints parallel to the surface of granite outcrops dividing the rock into sheets; (b) joints arranged concentrically; and (c) radial joints which converge towards the centre of the mass. The granite may thus be divided up into fairly widely-spaced blocks. The joints are additional to the faults and other fractures of tectonic origin.

The weathering phenomena occurring on granite surfaces disintegrate and decompose the minerals, particularly the feldspars. The resulting product is a fairly soft and even fairly friable layer of weathered granite of a much greater porosity and permeability than that of the unweathered rock. In certain climatic conditions the permeability of the weathered rock may be greater than that of the unweathered joint rock. In other conditions the forming of clayey material makes the weathered rock almost impermeable.

In the presence of water and CO_2, orthoclase weathers to kaolinite with the production of free silica, forming a layer of angular fragments on top of the less-weathered rock. This layer is generally permeable.

The thickness of the weathered zone depends on the climate, the duration of the weathering process and the intensity of the erosion. In mountainous regions there may be no weathered zone at all. On the other hand there may be a much larger number of open fissures, not only because of tectonic fracturing but also because of downhill gravitational movement on slopes. On rock surfaces which have been planed by ice action, only compact rock remains.

In temperate climates, the weathering zone may extend down to 30 m or more and water-bearing joints may reach depths of more than 100 m.

14.2.1.2 **Water-bearing granite**

Where there is sufficient permeability, the weathered and disintegrated granite contains ground water in a manner comparable with that in rocks having interstitial porosity.

At shallow depth, open sheet joints intersect the concentric and radial joints. If they contain water, the entire joint system may behave like an aquifer. With depth, sheet joints become increasingly widely spaced and less and less open, and the concentric and radial joints also close. In the granite of Massachusetts and Georgia, for example, the rock layers between joints are a few centimetres thick at a depth of a few metres (Le Grand, 1949).

The volumes of water decrease with depth as also does the intercommunication between joints. At greater depths the granite is virtually dry although some fractures may transmit water to a great depth, as has been observed in tunnels and mines.

Few measurements are available of the permeability of granite. The hydraulic conductivity of the unweathered rock is small, but that of weathered granite is somewhat greater: granites (Louis, 1968), 0.5 to 2.0×10^{-10} cm/s; weathered granite (Morris and Johnson, 1966) 3.7×10^{-4} to 1.8×10^{-3} cm/s from 8 samples.

For the hydraulic conductivity of a granite mass, Louis (1968) gives 0.7×10^{-1} cm/s in the direction of the fissures on the basis of one fissure 1 mm wide for every metre.

Lugeon's tests in Brazil (Franciss, 1970) gave $10^{-4}-10^{-6}$ cm/s with the Badushkin formula (about 100 measurements).

14.2.1.3 Springs, wells and boreholes in granite

Because of the more numerous fissures of granite in steep mountainous terrain, there are often more springs than in other rocks. Each spring has, however, only a relatively small discharge. In plains or hilly regions, the springs may emerge from the decomposed arkoses resting on the impermeable granite or from the well-fissured part of the unweathered rock.

Water is more abundant in low-lying areas, for the following possible reasons (Fig. 14.2.1.3): (a) owing to the relatively gentle gradients along the valley floors, runoff moves slowly and more of it infiltrates; (b) infiltration of runoff from the valley flanks accumulates beneath the valley floors; (c) seepage through the rocks and residual material of the slopes reaches the valley floors; (d) greater storage exists beneath the valley floors because of the thickness of the residual or alluvial deposits.

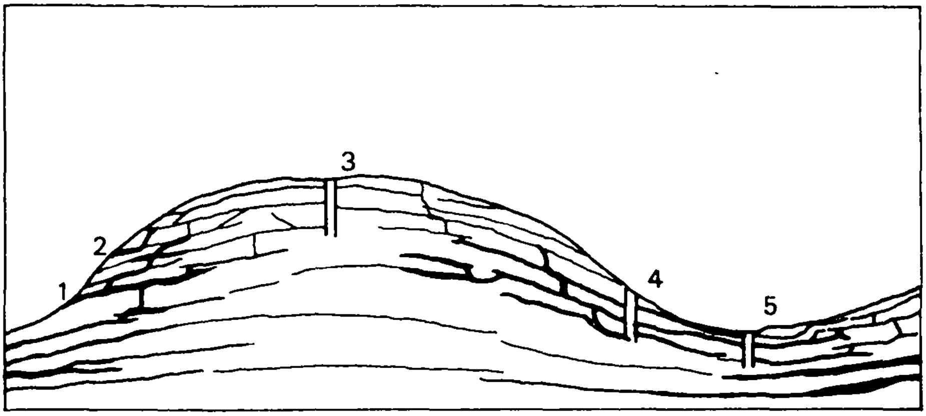

FIG. 14.2.1.3. Cross-section of sheeted terrain showing water-filled joints in heavy lines. 1. Perennial sheet spring; 2. Intermittent sheet spring; 3. Unsuccessful well; 4. Successful well; 5. Possible flowing well. (Legrand, 1949).

Wells should therefore be sited in those places where the weathered zone and the upper fractured zone of the granite are best developed. This means sitting wells on the flat valley floors because if they are placed on the flanks of valleys, particularly where these are steep, they may not be successful. Wells should not be deeper than about 100 m, as below this depth the chances of success are small.

Table 14.2.1.3a shows the yield of wells in Maine and their depths according to Clapp (1911). Table 14.2.1.3b gives the characteristics of wells in South Africa, according to Du Toit (1928), and some characteristics of wells in various regions are compared in Table 14.2.1.3c. In Sweden, the following average yields in litres per minute have been noted: Bohnsländ granite, 22; Smalande granite, 43; Filipstad granite, 32; Uppland granite, 23 (Meier and Peterson, 1951).

Generally speaking, a fair estimate would be that the average yield of wells ranges from 35 to 65 l/min. In intensively weathered or faulted rocks average yields may range from 150 to as much as 300 l/min. In Ukrainian crystalline rocks (Rudenko, 1958), the specific capacity of wells ranges between 11 and 16 l/min per metre of drawdown.

Yields of wells in granite are, generally speaking, small. Sheet joints, however, are sometimes the source of large supplies. Yield is also variable from one locality to another.

TABLE 14.2.1.3a. Yields and depths of wells in Maine

Yield (l/min)	Number of wells by depth					Number of wells
	15–30 m	30–60 m	60–90 m	90–120 m	120 m	
4.5–22.5	7	5	2	2		16
22.5–45	5	1				6
45–225	6	7	5			18
> 225	1	1	1			3
Yield unknown	19	6	3		1	29
Too little water	2	2	3		1	8
Water unusable			1	1		2
TOTAL	40	22	15	3	2	82
Success (%)	95	91	73	67	50	87

TABLE 14.2.1.3b. Characteristics of wells in South Africa

Old granite	Average				Average of yield (l/min)	Failure (%)
	Number of wells	Total depth (m)	Depth of water (m)	Depth to water surface (m)		
Pretoria–Johannesburg	62	40	21	11	84	14
North Transvaal	497	47	34	22	83	25
Rustenburg	130	61	51	40	65	50
South-West Transvaal	404	37	26	14	98	10
Mafeking	202	51	35	29	76	35
Vryburg	136	48	35	31	61	27
Van Rhynsdorp	16	56	43	35	16	31

TABLE 14.2.1.3c. Characteristics of wells in various regions

	Average depth (m)	Maximum depth (m)	Depth of water intake (m)	Average yield (l/min)	Maximum yield (l/min)
Connecticut (1)	27	91		48	
Uganda (2)	(63)		21–27	0.1–7.5	
			54–60	7.6–4.5	
Brazil (3)			30–40	25	310
Mauritania (4)				(11)	330
				30	47
Rajasthan (5)				(12)	

Sources: (1) Gregory and Ellis, 1909; (2) Gear, 1951; (3) Suszczynski, 1968; (4) Archambault, 1960; (5) Taylor, Roy and Sett, 1954.

Of two wells a short distance apart, one may be completely unsuccessful and the other may provide an acceptable or even a large yield.

Geological study is needed to find out why some wells yield more than 30 l/min and why some are better than others. It is necessary to know the spacing of joints and how open they are, as well as the existence of faults and other structures and their geometry, and how much water they can transmit. Only where the granite is exposed can all this be done. Otherwise, geophysical methods can be helpful in conjunction with drillings.

The best chance of finding water is where the granite is covered by a thick layer of permeable weathered material, especially if this overlies a depression in the unweathered granite surface.

14.2.2 *The hydrogeology of metamorphic rocks*

The rocks considered here are those comprising the large, ancient massifs, and shields resulting from a regional metamorphism. Examples include crystalline schists, slates, amphibolites, quartzites and crystalline limestones.

14.2.2.1 Fissures and fractures

Some characteristics of gneiss, crystalline schists and slates include: (a) cleavage, particularly in slatey rocks; (b) schistosity, i.e. parallel layers of flakes having the same petrographic composition and independent of the stratification; and (c) foliation, i.e. parallel layers of flakes of different mineralogical composition and independent of the bedding. Micaceous minerals are most often found in the planes of foliation. In some cases the bedding planes have been preserved.

Additional to these partings are the sheet joints, the joint systems roughly perpendicular to the fractures mentioned above, and fissures enlarged by weathering and faults.

14.2.2.2 Intensively metamorphosed rocks

Water-bearing openings, aquifers. Partings due to schistosity and foliation and sheet joints, where they exist, have insignificant width as compared with those in granite. Although they may be closed at shallower depth than in granite, they may play a considerable role in ground-water storage. By reason of the steep dips of the closely spaced partings in schists and slates, the ground surface intersects them at a wide angle. Where these partings are enlarged by weathering, small increments of infiltration may take place over wide areas. This infiltration may then move downward and enter other fissures and fractures to provide limited ground-water supplies.

Sheet joints result from decompression which occurs when erosion has removed a considerable thickness of rock. They may play a role in the hydrogeology of certain metamorphic rock such as gneisses. In the Archaean shield of south-western Australia, for example, joints developed in this way are subparallel to the existing natural surface and are often very open, as much as 10 cm in gneisses. These extensive sheet joints occur one beneath another in the solid rock to depths of 100 m or more. The deep joints are less open than the shallow joints because of increasing pressure with depth. The shallow joints are known to be capable of storing and allowing the movement of ground water to depths of at least 25 m. Bore yields of 40 m^3/h (666 l/min) are not uncommon.

Sheet joints in gneisses also exercise a marked control over the depth of the weathered zone. This in itself is important, for two reasons. In many cases a massive crystalline metamorphic rock will weather into a moderately permeable material that will allow the downward percolation of surface water, which can be recovered by wells or bores put down to the hard rock, if the topography is favourable and the weathered mantle is thick enough. The thickness needed for an effective aquifer depends, among other things, on the topographic relief. In valley floors with flat gradients, possibly 5 to 10 m is sufficient. On long hillslopes of 10° and 30°, a depth of 20 to 40 m may be needed, because otherwise the ground-water movement downslope may deplete the volume of stored water during the dry season. In semi-arid areas, if the weathered mantle is thicker than about 50 m, a saturated zone may not exist, because infiltrating water is stored as soil moisture and then dissipated. In these conditions, drilling or well-sinking to greater depths cannot be recommended.

The yields from shallow wells and intensively metamorphosed rocks such as gneiss may be substantial and sustained if they tap open fissures and fractures which drain large areas. With increasing depth, the fissures and fractures close more rapidly than in granite. Being unidirectional, they are less interconnected, and yield smaller and more easily depleted amounts of water than are found in granites at similar depth.

Permeability and yield. The porosity and permeability of rocks themselves are very low. The porosity of gneiss is about 0.1 to 3 per cent and that of schists from 0.5 to 5 per cent. Depending on the degree of weathering, the porosity in both rock types may increase to 50 per cent. The average from eighteen samples examined at the hydrological laboratory in Denver, Colorado, was 38 per cent (extremes 4.4–49.3 per cent) (Morris and Johnson, 1966). The hydraulic conductivity of the rock itself is extremely low, as shown in Table 14.2.2.2a.

TABLE 14.2.2.2a. Hydraulic conductivity in different types of metamorphic rocks

Rocks	Number of samples	Hydraulic conductivity (cm/s)			Region
		Minimum	Maximum	Average	
Slates (1)	8	2.1×10^{-10}	5.3×10^{-8}	1.0×10^{-8}	United States
Schists (1)	17	2.1×10^{-9}	1.3×10^{-3}	2.1×10^{-4}	United States
Schists (2)		0.7×10^{-9}	1.2×10^{-7}		
Slates (3)	9	5.0×10^{-10}	4.5×10^{-8}	1.3×10^{-9}	Marquette (Michigan)
Micaschists (3)	1			6×10^{-9}	Marquette
				2.1×10^{-9}	Marquette
Metamorphic rocks (4)			1.0×10^{-5}		Michigan
Gneiss (5)	34 tests	8.0×10^{-4}	2.0×10^{-7}		Brazil
Granite-gneiss (5)	72 tests	4.0×10^{-4}	5.0×10^{-7}		Brazil
Slates (5)		3.0×10^{-4}	5.0×10^{-7}		Brazil
Schists (6)		5.4×10^{-5}	1.6×10^{-4}		Ukraine
Ferruginous schists (6)		7.1×10^{-4}	4.2×10^{-3}		Ukraine

Sources: (1) Morris and Johnson, 1966; (2) Louis, 1968; (3) Stuart, Brown and Rhodehamel, 1954; (4) Walton 1962; (5) Franciss, 1970; (6) Rudenko, 1958.

Pumping tests on fissured rock massifs give values which may be sometimes 1,000 times greater than the hydraulic conductivity of the rock itself. The hydraulic conductivity calculated by Louis (1968) was 0.6 cm/s in the direction of the fractures, assuming one fracture 2 mm wide for every metre in a mass of schists.

Permeability along the direction of fracturing is two to three times greater than the average. Thus, the permeability of metamorphic rocks is highly directional.

As the rock blocks between the joints are relatively large, it is not really correct to speak of anisotropic permeability, except in the case of very large rock volumes. On a smaller scale, it is preferable to speak of directional permeability (Vecchioli, 1967).

The systematic studies carried out in metamorphic crystalline terrains of north-east Brazil have shown that fractures crossing obliquely or parallel to the structural directions of the rocks have different water-yield capacities. Fractures perpendicular to the schistosity or cleavage give more water than oblique fractures, and oblique fractures more water than parallel fractures (Suszczynski).

In the Monteiro region of Brazil (Suszczynski, 1968), the mean yield is thus: 5.5 m^3/h for twenty boreholes in perpendicular fractures; 2.8 m^3/h for twenty boreholes in oblique fractures; 1.4 m^3/h for twenty boreholes in parallel fractures.

In the Brunswick shales in New Jersey, Vecchioli (1967) determined the transmissivity (T) parallel to the direction of the beds and perpendicularly, as well as the storage coefficients (S) (Table 14.2.2.2b). Well yields in Connecticut are shown in Table 14.2.2.2c (after Gregory and Ellis, 1909), while Table 14.2.2.2d gives mean yields obtained from other sources.

TABLE 14.2.2.2b. Transmissivity (T) and storage coefficients (S) in the Brunswick shales, New Jersey (Vecchioli, 1967)

Direction of bed	Well	T (m²/s)	S
Perpendicular	Pumped well	1.44×10^{-2}	
	Well No. 1	1.27×10^{-2}	2.49×10^{-4}
	Well No. 2	1.03×10^{-2}	2.51×10^{-4}
Parallel	Well No. 3	0.82×10^{-2}	2.51×10^{-4}
	Well No. 4	0.82×10^{-2}	2.51×10^{-4}

TABLE 14.2.2.2c. Well yields in Connecticut (after Gregory and Ellis, 1909)

	Number of wells	Thickness of overburden (m)	Depth into rock (m)	Total depth (m)	Yields (m^3/h)[1]
Gneiss	70	5	35	40	2.79 (50)
Quartzite schists	3	10.3	125	135	1.64 (3)
Schists	23	4.2	23	33.4	4.16 (16)
Slate	5	4.4	5	28.8	Very small (5)

1. The figures in parentheses indicate the number of wells in which the yield has been measured.

TABLE 14.2.2.2d. Mean yields in different types of metamorphic rocks

Country	Type of rock	Yield (m^3/h)	Country	Type of rock	Yield (m^3/h)
Sweden (1)	Hallan gneiss	2.76	Brazil (3)	Monteiro gneiss[1]	4.62
Sweden (1)	Västergötland gneiss	3.18	Brazil (3)	Petrolina gneiss[2]	2.16
Sweden (1)	Bohnslan gneiss	1.08	United States (2)	Maryland schist	5.46
Sweden (1)	Värmland gneiss	2.40	United States (2)	Virginia schist	2.70
Sweden (1)	Södermaland gneiss	0.72	United States (2)	New England schist	2.28
Sweden (1)	Norrland migmatite	2.52	Brazil (3)	—[3]	1.50
United States (2)	Maryland gneiss	2.70	India (4)	Rajasthan slates[4]	2.22
United States (2)	Virginia gneiss	3.84	United States (4)	Maine slates	3.42
United States (2)	New England gneiss	2.04			

1. Thirty-three boreholes.
2. Forty-six boreholes.
3. Twenty-five boreholes.
4. Twenty-five wells with discharge ranging from 21 to 47 l/min.

Sources: (1) Meier and Peterson, 1951; (2) Davis and de Wiest, 1966; (3) Suszczynski, 1968; (4) Taylor, Roy and Sett, 1954.

14.2.2.3 Quartzite, amphibolite and crystalline limestone

The best aquifers among metamorphic rocks are quartzite and marble, one because it is competent and both develops and preserves a good joint system; the other because of its solubility and the likelihood of openings developing along the joint planes. Where the beds of quartzite or marble are not thick and are enclosed in incompetent rocks, such as schists and phyllites, these rocks may act as drains because of the large number of joints.

14.2.3 *Non-carbonate, indurated, sedimentary rocks*

14.2.3.1 Sandstones

Sandstones are porous both because of their interstices and because of their fissures. They are thus transitional between the unconsolidated and compact indurated rocks. The variable interstitial porosity, ranging from less than 1 per cent to about 35 per cent, depends not only on non-uniformity of grain sizes and the way they are packed but also on their calcareous or siliceous cement. Fissure porosity varies greatly. Joints are often abundant and are more open when the sandstone is indurated. Bedding planes also may increase the porosity.

There are two kinds of permeability: through intergranular openings, and through fissures. Because of varying degrees of obstruction caused by materials, interstitial permeability varies as shown in Table 14.2.3.1a.

TABLE 14.2.3.1a. Hydraulic conductivity and porosity in sandstones

Rock	*Hydraulic conductivity (cm/s)*				*Porosity (%)*			
	Minimum	Maximum	Average	Number of samples	Minimum	Maximum	Average	Number of samples
Fine sandstone (1)[1]	4.2×10^{-7}	1.9×10^{-3}	2.6×10^{-4}	20	13.7	49.7	33	55
Medium sandstone (1)	2.6×10^{-6}	1.2×10^{-2}	4.1×10^{-3}	13	29.7	43.6	37	10
Miocene (Tunisia) (2)	1×10^{-3}	1×10^{-2}			20	37		2
Oligocene (Tunisia) (2)	3×10^{-3}	5×10^{-3}						
Oklahoma sandstone (3)	1×10^{-7}	5×10^{-5}						
Devonian sandstone Bradford (3)			2.6×10^{-6}				14.8	
Mississipian Berea (3)			3.5×10^{-4}				19	
Ordovician, Oil Creek (3)			3.9×10^{-6}				6.7	
Cretaceous Woodbine (3)			4.3×10^{-3}				25.6	
Pliocene Repetto (3)			3.5×10^{-5}				19.1	
Eocene Wilcox (3)			2.9×10^{-7}				15.3	
Carboniferous (4)	2.9×10^{-12}	6×10^{-11}						
Devonian (4)	2.1×10^{-11}	2×10^{-11}						

1. Refers to horizontal hydraulic conductivity. The vertical hydraulic conductivity in cm/s taken over twenty samples lies between 5.3×10^{-7} and 2.5×10^{-3}, with an average of 3.7×10^{-4}.

Sources: (1) Morris and Johnson, 1966; (2) Schoeller, 1962; (3) Davis and de Wiest, 1966; (4) Louis, 1968.

The permeability of fissures or of rock mass is difficult to measure. The specific capacity of wells, however, may be used to estimate permeability. Louis (1968) gives the following figures for hydraulic conductivity along fissure directions in sandstones of various ages at 1 m spacings: 0.2 mm fissure, 0.6×10^{-3} cm/s; 0.4 mm fissure, 0.5×10^{-2} cm/s; and 0.7 mm fissure, 2.5×10^{-2} cm/s.

In the Ukraine (Rudenko, 1958) the permeability of quartzites varies between 2.3×10^{-4} and 1.1×10^{-3} cm/s. In Tunisia, Schoeller (1939) and Gosselin and Schoeller (1939) obtained approximate values by pumping from boreholes (Table 14.2.3.1b).

TABLE 14.2.3.1b. Hydraulic conductivity in Tunisian sandstones

Hydraulic conductivity (cm/s)	Number of boreholes		Hydraulic conductivity (cm/s)	Number of boreholes	
	Middle Miocene sandstones	Oligocene sandstones		Middle Miocene sandstones	Oligocene sandstones
1×10^{-4}–1×10^{-3}	7	8	2×10^{-3}–3×10^{-3}	3	2
1×10^{-3}–2×10^{-3}	4	3	3×10^{-3}–7×10^{-3}	4	3

In the middle Bunter sandstones of the Saar where there are 23 per cent of NNE-SSW fissures, 14 per cent NE-SW, 15 per cent WNW-ESE, 14 per cent NW-SE and 20 per cent NNW-SSE, Seiler (1969) gives the specific capacity which can be as high as 43.2 m³/h/m, the most frequent specific capacities ranging between 1.8 and 5.4 m³/h/m. The fractures may receive infiltration from the surface and they may also obtain water from the sandstone itself, particularly where it is porous.

When water is pumped from a well or a borehole in sandstone, two types of permeability are involved. At the start of pumping, water will be drawn both from fractures and intergranular spaces. Water will move more freely through the fractures. Yield will be more effectively sustained where the fractures are numerous and extensive. During the later stages, the draining function will predominate and the permeability of the rock itself will control the yield.

The yield from sandstone does not increase in proportion to the depth. Below a certain depth, as for example 100 to 200 m in average Bunter sandstone, the specific capacity tends to remain constant because fracturing also diminishes with depth.

Finally, Rats and Chernyashov (1967) demonstrated a statistical relationship between hydraulic conductivity, fissuring and thickness of the rock layers, at least in the sandstones of the Ordovician flysch in central Kazakhstan:

$$\log l = 0.41 \log m + 0.45; \quad \log \Delta l = 0.41 \log m - 0.61; \quad K_f = 0.82 \log m - 0.19$$

where:

l = average distance between neighbouring fractures, in metres;
Δl = average width of beds, in metres;
m = thickness in metres;
K_f = hydraulic conductivity in metres per day, along the fractures perpendicular to the bedding.

Permeability is thus not directly proportional to the thickness of the beds.

14.2.3.2 Shales

Shales have low porosity and water can circulate only in the fissures (see Table 14.2.3.2a). Gondouin and Scala (1958) give the following figures for hydraulic conductivity in shales: Pennsylvanian (approximate depth 150 m), 9×10^{-11} cm/s; Cretaceous, 4×10^{-9} cm/s.

Shales are poor aquifers but in some places, in the absence of good aquifers, wells in shale may yield some water from joints and bedding planes. In general, the joints are best developed near the surface, and are better developed in the more indurated shales. In places, it may be possible to obtain usable quantities of water where considerable fissuring and weathering exist.

TABLE 14.2.3.2a. Porosity in shales

Shale	Porosity (%)			Shale	Porosity (%)		
	Minimum	Maximum	Average		Minimum	Maximum	Average
United States (1)	1.4	9.7	6.4	Chanute shale (Pennsylvanian) (2)			15
Gros Ventre formation (Cambrian) (2)			11.1	Nonesuch shale (Siliceous) (2)			1.6
Graneros shale (Cretaceous) (2)			24.9				

Sources: (1) Morris and Johnson, 1966; (2) Manger, 1963.

14.2.4 *Extrusive volcanic rocks*

Those volcanic rocks, deficient in silica and high in iron and magnesia, tend to be the best aquifers. The silica-deficient basalts, for example, are often good aquifers whereas the silica-rich rhyolites and trachytes seldom contain large quantities of ground water.

14.2.4.1 **Basalts**

Basalts are commonly among the most productive of all fractured-rock formations because of the numerous openings they contain. Moreover, they are widespread as shown as follows: 500,000 km^2 in India, the 'trap rocks' of the Deccan; 650,000 km^2 in North America; 900,000 km^2 in Brazil; 100,000 km^2 in Ireland and the United Kingdom; etc. They therefore represent extensive sources of ground water.

The characteristics of basalts depend on the nature of the flows: *aa* lavas, with a hardened, scoriaceous crust due to the escape of gases; pahoehoe lavas, with smooth, undulating ropy surfaces; pillow lavas, formed by outflows under water, the pulverised material being sometimes permeable.

Flows may be superimposed on each other, varying in thickness and number. Compact basalt also alternates with scoriaceous crusts or ropy surfaces. There may be a layer of ash between the flows or even a layer of fossil soil, alluvials or other debris if the dormant phases during extrusions were fairly prolonged. Layers of ash do not commonly occur in pahoehoe flows.

Openings and permeability. The productivity of basalts is due to the numerous possible openings. These, in decreasing order of importance, are: (a) scoriae; (b) breccia zones between *aa* flows; (c) cavities between pahoehoe lava flows; (d) shrinkage cracks, which may be of two kinds—one perpendicular to the isothermic surfaces which have been caused by surface coolings, the other caused by cooling from beneath and which cut through the basalt flow in prismatic columns; (e) gas vesicles; (f) lava tubes within flows, particularly in pahoehoe lavas; (g) fractures, the result of mechanical forces acting on the cooled lava; (h) tree moulds.

Horizontal permeability is much greater than vertical, because shrinkage cracks, scoriaceous, brecciated and cavernous zones are concentrated between flows. The internal parts of the flows are generally impermeable. Vertical permeability is sometimes so slight that confined aquifers may be formed, giving pressure differences of tens of metres.

Generally, pahoehoe flows are more permeable than the thick and dense *aa* flows. The lower and upper parts of *aa* flows may be brecciated and reasonably permeable. Mesozoic and Palaeozoic lava flows are generally less permeable than flows of younger age because the openings have been infilled. In pillow lavas, water can circulate between the pillows if not blocked by vitreous material or secondary minerals.

Permeability values. Non-vesicular basalt has a low porosity. Even with an abundance of vesicles, the permeability of the rock itself is low (Table 14.2.4.1a).

TABLE 14.2.4.1a. Hydraulic conductivities and porosities of basalts in the United States (Morris and Johnson, 1966)

	Minimum	Maximum	Average	Number of samples
Hydraulic conductivity (cm/s)	2.1×10^{-9}	4.7×10^{-5}	1.05×10^{-5}	93
Porosity (%)	3	35	17	94

There are few hydraulic conductivity values (K) for flows or series of flows. Transmissivities (T) and conductivities (K) for the basalt flows of the Oudja region in Morocco are given in Table 14.2.4.1b. The storage coefficients range between 2×10^{-2} and 7×10^{-2}.

According to Franciss (1970) hydraulic conductivity measured by Lugeon tests and calculated according to the Babushkin formula is 3×10^{-4} to 5×10^{-7} cm/s (156 measurements) for basalts in Brazil.

TABLE 14.2.4.1b. Transmissivities and hydraulic conductivities of the basalt flow of the Oudja region (Mortier, Quang and Sadek, 1967)

	K (cm/s)	T (m²/s)
Weathered basalts and fissured compacted ash	1.7×10^{-2} to 2×10^{-1}	1.1×10^{-2} to 3×10^{-2}
Basaltic tuffs and slightly fissured compact basalts	1.2×10^{-1} to 7.5×10^{-1}	1.3×10^{-5} to 7.5×10^{-3}

Ground water. Basalt flows may cover large, mainly horizontal or slightly inclined areas and may be superimposed on each other to considerable depths. The underlying topography may also be irregular. A flow may follow a valley for a considerable distance just as would a river, either as one or several flows. Numerous superimposed flows may, finally, constitute shields, as in the Hawaiian Islands.

Lava plains with a great number of superimposed flows may hold a vast quantity of water, as in the Snake River Plain (Idaho) of the United States. Because of high permeability, the potentiometric surface has a low gradient and in places is as deep as 250 m beneath the surface. For the same reason, the unconfined ground-water body is thin and in places interrupted by highs on the basement surfaces. In places, depressions and burried valleys will act as drains. Water circulates through the basalts not only in fissures but also in scoriaceous layers, in ash and alluvium between the basalt flows, particularly in alluvium of buried valleys. The ground water is recharged naturally from the surface and also from rivers passing underground at the edges of basalt flows. For a distance of 320 km, all rivers coming down from the mountain disappear at the northern edge of the Snake River Plain. The water reappears in springs along canyon walls from permeable zones beneath or between flows.

Valley-filling basalts have simpler characteristics, generally with a small number of superimposed outflows. Ground water occurs either at the base of each flow or in the permeable layers, alluvials or ash. The inclination of the potentiometric gradient corresponds to the slope of the substratum. Recharge takes place directly from the surface, from the upstream end of the flow and from sheet runoff or streams at the edges of the flow. Springs occur at the ends of flows or else at breaks in slope on the floor of the substratum.

Lava shields, like those in the Hawaiian Islands, may contain bodies of fresh water to depths of as much as 350 m. At Honolulu on the island of Oahu, the basal fresh ground water rests on top of salt water (Stearns and Vaksvik, 1935). In addition perched water accumulates on sills, ash, consolidated tuffs, impervious alluvials or compact layers of basalt.

Recharge. Direct recharge from rain on basalts occurs readily where there is no vegetative cover, as in arid regions. The rainwater runs over the bare rock and seeps down into the fissures. Where vegetation covers the surface much potential recharge is lost through evapotranspiration.

On the Hawaiian island of Maui, 3.6 m of rain infiltrated out of a total of 5 m. On the island of Oahu, the inflow amounts to about 25 per cent of the annual rainfall (Stearns and Vaksvik, 1939) which varies from 500 to 6,350 mm.

In the Territory of the Afars and the Issas (Pouchan, 1973)[1] in a hot and arid climate, the recharge of ground water in basalt is about 10 per cent of the annual rainfall which is between 100 and 250 mm. There is virtually no vegetation.

In short, indirect recharge from streams disappearing into basalts can be important.

Springs. Springs issuing from basalts are among the largest known and compare in importance with those rising from carbonate rocks. One group of basalt springs in California and Oregon has a combined discharge of 40 m^3/s, of which the Datta spring alone supplies 1.4 to 3 m^3/s.

In Idaho, the basalt springs along the canyon walls between Milner and King Davis discharge about 110 m^3/s, each of the springs yielding 4 to 16 m^3/s. The famous Thousand Springs, which rise between the basalt flows, yield 15 to 20 m^3/s (Meinzer, 1927).

At Pearl Harbor, on Oahu island, the *K*alauaa spring yields between 0.4 and 1 m^3/s and the Kaluaoopu spring between 0.7 and 1 m^3/s (Stearns and Vaksvik, 1935).

The flow from the springs may be relatively constant where it is sustained by a large body of ground water and where the fissures are narrow and ash is present. On the other hand, flow may be variable if wide fissures are well developed and abundant.

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[1] Private communication.

Wells and boreholes. The individual yields from wells and boreholes in basalt can differ greatly from point to point. Yields of from 5 to 8 l/s are not uncommon and some may exceed 60 l/s. In Honolulu, wells produce 25 l/s on the average with drawdowns of only 0.3 to 2 m. One well yields 170 l/s. In the Territory of the Afars and the Issas boreholes in basalts give 1 to 10 l/s and 2 l/s in weathered basalts (Pouchan, 1973). In general, the specific capacities range between 0.7 and 4 l/s/m. Some boreholes, however, yield as little as 0.1 l/s/m and others as much as 33 l/s/m.

In the Yakkima basalt of the Dalles (Oregon) area, the lower water-bearing zone has specific capacities ranging between 7 and 83 l/s/m (Piper, 1932). Foxworthy and Bryant (1967) mention specific capacities of between 12 and 26 l/s/m in the basalts of the Dalles.

In the prismoidal *aa* basalts of the Territory of the Afars and the Issas, Pouchan (1973) reports specific capacities of 13, 20, 32, 52 and 72 l/s/m, but only 0.1 l/s/m in a weathered basalt.

The yield of wells depends on the permeability of the rock, as discussed above. The regional water-table may lie at considerable depth and perched water bodies commonly occur at shallow depths.

Tunnels. Tunnels have been driven especially to tap perched water on sills or on impermeable layers between flows. On Oahu, there are yields of 12 l/s in a gallery 60 m long and 547 l/s in a gallery 660 m long (Waiahole Kahoua 1) (Stearns and Vaksvik, 1939).

Dykes. Dykes frequently form barriers to horizontal flow and consequently bodies of ground water form between them or against one wall.

Heat from the injected mass of dyke rock tends to bake the country rock, causing rapid heating and cooling which tends to fracture the rock. The fractures store water, and drilling into the country rock close to a dyke may be worthwhile. In the Precambrian crystalline basement of north-west Africa, in western Australia and in north-east Brazil, fractured basic dyke systems are explored for ground water. Under certain climatological conditions, dykes of quartz and pegmatite are sufficiently fractured to contain valuable aquifers.

14.2.4.2 Trachytes and rhyolites

These rocks are usually more massive than basalts. In addition, there are fewer and narrower fissures and the brecciated parts are compact. They thus have a relatively low permeability and are not so valuable as a water resource. As a general rule, wells in these rocks provide only a small amount of water, and the springs have a small yield. In exceptional circumstances, however, some springs may yield large volumes, as much as 5 m^3/s, particularly when there are large faults, such as in the Nandewar and Warrumbungle Ranges of eastern Australia.

14.2.4.3 Extrusive features

These include pipes, necks and calderas covering much smaller areas than do the great basalt flows.

Calderas exhibit a system of circular fractures in which the ground water may have different potentiometric levels that become lower and lower towards the centre. In the volcanic islands of Oceania and in African volcanic ranges, these structures are well developed and contain ground water.

Pipes and necks generally have divergent systems of radial fracture. There may still be concentric fractures but they dip in the opposite direction to those of calderas having internal subsidence.

14.2.4.4 **Prospecting**

Prospecting for ground water in such fissured and fractured rocks requires, first, a much more thorough geological study than that in sedimentary terrains, and also a complete inventory of all wells and springs. Lava flows must also be studied in detail. Geophysical techniques may be difficult to apply, except where basalt lies directly on the rock having different physical characteristics.

14.2.5 *The quality of water from fissured, non-carbonate rocks*

The silicates present in all the fissured non-carbonate rocks give the waters a particular chemical composition.

As the process of dissolving the silicates from the rocks is slow, the water will have only a low ion content under the usual conditions in temperate zones, ranging from a few dozen to a few hundred mg/l.

Because of the non-carbonate nature of the rocks, the HCO_3^- and CO_3^{--} ions will come only from the breakdown of the CO_2 dissolved in the water. They are consequently found in quantities roughly half those in ground water which is in contact with limestones, that is from 50 to 150 mg/l of HCO_3^-.

The content of Cl^- originating from the rock is also minimal because only traces are contained in minerals such as chlorapatite. A relatively large proportion comes from rainwater. Cl^- content thus depends in large measure on the concentration brought about by evapotranspiration.

The SO_4^{--} content depends mainly on the oxidization of sulphides, particularly pyrites. The rSO_4/rCl ratio is thus variable.[1]

The common cations Ca, Na, Mg and Fe originate mainly from the rock itself, but the ionic dissociation from the silicates is difficult. Their relative proportion depends on the nature of these silicates. Rocks rich in pyrites give large iron concentrations.

As there are many more Na and K ions than Cl^- in the rocks, there must therefore be an inherent alkali to chloride disequilibrium $rCl - r\,(Na + K) < 0$, which should not be interpreted as a base exchange. As there are fewer HCO_3^- ions in relation to the normal dissolved CO_2 content in ground water, the $pH = \log 1/K_1 + \log \langle HCO_3^- \rangle / \langle CO_2 \rangle$ is low, at the start from 4 to 6. After the water has been exposed to air, however, allowing CO_2 to escape, the pH may become alkaline and rise to above 8. Oxidization of pyrites produces sulphuric acid and lowers the pH.

The $rCa/(rNa + r\,K)$ ratio for granites and gneisses depends essentially on the nature of the feldspars. In water originating from basalt, this ratio is most frequently higher, as is the rMg/rCa ratio. The silica content is also greater.

Water from schists and slates usually contains a higher proportion of all salts, mainly because of more active evapotranspiration resulting from slower downward movement. As these rocks frequently contain pyrites, iron contents are high.

When sandstone and quartzite are entirely siliceous, they contain water with a mineral content even lower than that of granite and gneiss because the input of cations is very small.

The chemical composition of the water also depends on climatic conditions which affect the vegetative cover and consequently evapotranspiration. The higher concentration of all salts in a hot and arid region is essentially the result of the greater evapotranspiration affecting a small quantity of water which has seeped into the ground.

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[1] r = milliequivalent.

As one example, in north-east Brazil the average dry residue is 1.664 to 8.038 g/l for a rainfall of 350 to 624 mm/year, but when infiltration is rapid some well waters contain only 0.360 to 0.460 g/l of dissolved salts (Suszczynski, 1968).

14.2.6 *The influence of the climate, the environment and the relief*

The extent of each fracture system exposed at the surface, the extent to which the spaces in it are filled, and the nature of the weathering zone that covers them depend largely on external geological processes and primarily on the environment.

The peneplaned crystalline basement rock is almost devoid of vegetation in semi-arid north-east Brazil. This hydrogeology contrasts with that of a similar geologic terrain in the extremely humid tropical region of Amazonia where there is a thick vegetative cover.

The systems of regional faults or fractures along and across mountain ranges are generally covered by some tens of metres of weathred material. This material is saturated with permanent ground water which feeds the underlying fractured systems. In this humid region, plant assemblages are an indication of buried water-bearing fractures beneath the weathered materials.

14.2.7 *Geological and hydrological criteria for siting of boreholes*

The drilling of boreholes for hydrogeological exploration and the extraction of ground water should take into account several criteria. The hydrogeology of fissured and fractured rocks is defined in sections 14.2.1 to 14.2.4. Particular note should be taken of the hydrogeological significance of generic types of fissures, the decrease in permeability with depth, the role of weathering and non-geological features such as regional hydrology (precipitation, surface runoff, infiltration), relief, accessibility for drilling, and distance from places where the water is to be used.

Because water-bearing rock and underground circulation vary greatly, permeable and impermeable rocks should be identified.

Drilling should be concentrated in areas of maximum fissuring and fracturing. Systematic measurements in the field should be made of the dip and direction of all fissures and fractures. A knowledge of the fracture systems, both local and regional, is important because they often form the most active conduits for water circulation.

Heavily fractured zones constitute the best places for drilling. Attention should be paid to the mineralogical composition of the rocks. Formations likely to contain unusable water should be avoided. In arid and semi-arid regions the relation between fracture systems and streams is important. A stream may replenish the water-bearing rock by infiltrating its fractures, especially the transverse ones. Sites for drilling should be chosen so as to penetrate fracture systems which have a good hydraulic connexion with a stream. The size of the stream and the nature of its alluvium are also important. Boreholes should be located as close to the stream as possible without incurring the risk of flood damage.

14.3 Hydrogeology of carbonate rock terrains

14.3.1 *Introduction*

Carbonate rocks may contain large volumes of water, often greater than those of terrains with interstitial permeability or fissured terrains, because of high storage coefficients. Also, loss by evapotranspiration may be low and surface runoff almost non-existent. In many cases, the water is of better quality than that in other kinds of rocks. The $CaCO_3$ in pure limestones has a low solubility limit and, if the rock has no impurities and evapotranspiration is low, so is salinity.

The hydrogeology of limestone terrains has a special character that requires separate treatment. Among the published works available are Cvijic (1960), Fenelon (1967), Herak and Stringfield (1971), Lehmann (1932), Maksimovic (1969), Sokolov (1967). The hydrology of carbonate rocks is also fully discussed in the *Guide to the Hydrology of Carbonate Rocks* prepared by the IHD working group on hydrology of carbonate rocks of the Mediterranean area (in press).

Speleology will not be discussed in the present work.

14.3.2 *Characteristic openings in carbonate rocks*

14.3.2.1 **Different rock types**

It is important to distinguish the characteristics of carbonate rocks according to their behaviour in karst hydrology.

Massive limestones and dolomites are thick and compact, with little interstitial porosity and with widely spaced joints and fractures. Ground-water flow is concentrated in widely spaced solution channels which are poorly interconnected.

Porous limestones, of varying thickness, including calcirudites, calcarenites and chalks, contain water which circulates with more or less difficulty depending on the size of the pores. Bedding planes and joints are more numerous. Circulation channels may occur.

Compact fine-grained bedded limestones and dolomites in large masses, including the lithographic or semi-lithographic limestones, have low porosity. Water circulates initially through numerous bedding planes and joints. These become enlarged and circulation increases accordingly.

Bedded marly limestones, alternating with marls, usually form poor aquifers.

14.3.2.2 **Types of openings**

Interstitial porosity. Primary porosity is the residual porosity of the sediment from which the rock was formed. It is frequently reduced by the deposition of calcite either during the process of lithogenesis or subsequently.

Secondary porosity is produced, in particular, by the dissolving action of circulating water.

Interstitial porosity is extremely variable. In certain calcirudites and calcarenites and chalk, porosity ranges from 1 to 20 per cent, and even up to 66 per cent for tufa.

Fissure porosity. In limestones, fissure porosity occurs essentially in fractures, joints and bedding planes. Expansion fissures due to removal of overburden are important, as are also exfoliation cracks near a vertical face or steep slope, faults and fault crushed zones (Fig. 14.3.2.2a).

Cavities. Circulating water within the limestone masses causes cavity porosity. Cavities are enlarged and extended upstream, creating directional permeability.

Several types of cavities can be distinguished:

Cavities in the zone of aeration or infiltration (Figs. 14.3.2.2b–f). Joints and fractures tend to become enlarged in this zone, particularly where they intersect. This effect tends to decrease with depth. Chimneys are vertical cavities opening downward at the end of a gallery or upward from a gallery or chamber. Sinkholes or shafts are vertical cavities emerging at the surface. 'Estavelles' are sinkholes into or out of which water flows, depending on the ground-water level. Passages, galleries or chambers form below the water-table.

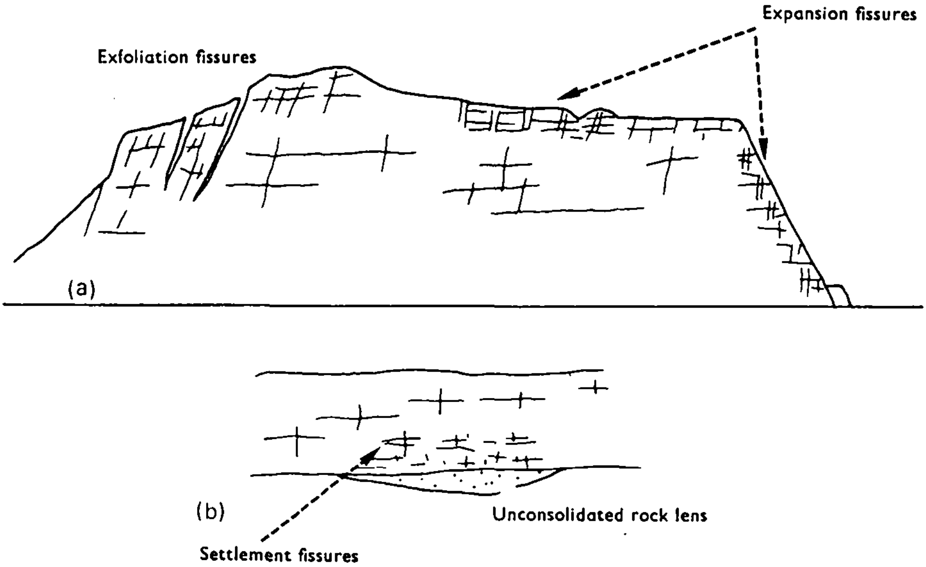

FIG. 14.3.2.2a. Types of fissures: (a) joints in the central part of the massif, expansion fissures, close to recently unloaded surfaces and exfoliation cracks in the faceward drainage zones near steep outward faces; (b) in the central part of a massif, fissures caused by differential settlement. (Renault, 1967.)

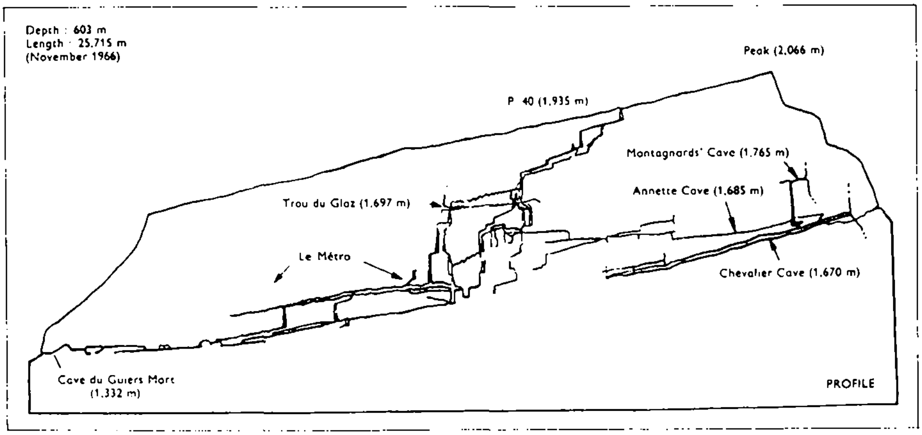

FIG. 14.3.2.2b. Caves of the Dent de Crolles.

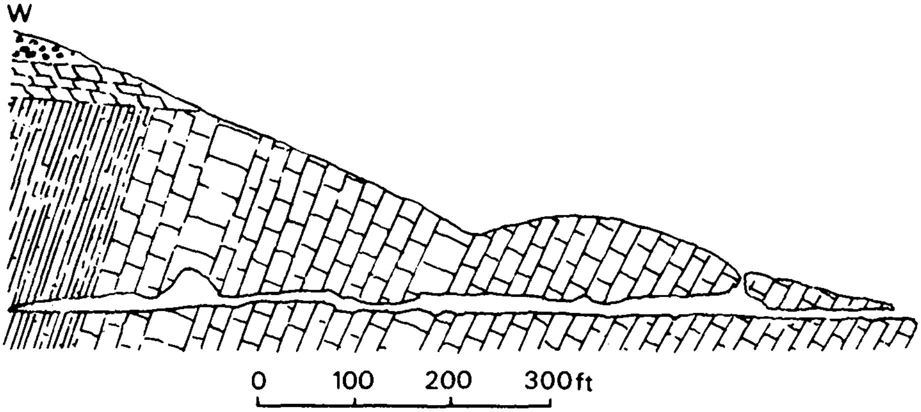

FIG. 14.3.2.2c. Lehman Caves, Nevada. Many caves are nearly horizontal, as shown in this profile. These caves are thought to have formed directly below a horizontal water-table. (Moore and Nicolas, 1964.)

Cavities in the saturated zone. Usually the gently sloping passages are single, labyrinthine or dendritic. These passages may be of primary or secondary origin (Renault, 1967, 1968, 1970). When the flow-rate is low they may be partly filled with sediments; when the flow rate exceeds 10 cm/s all the sediment is removed. Chambers are merely enlarged passages, sometimes with a domed roof (dome). Channels are pipe-like conduits.

These cavities occur mostly in the upper part of the zone of saturation. In the lower parts of the saturated zone, the rate of water movement is low, and the fissures or channels are much less subject to enlargement. They therefore are usually smaller than the ones higher up.

14.3.3 *Natural controls of circulation*

14.3.3.1 **The different zones of flow**

In some regions, as for example in Florida, Puerto Rico, the Aquitaine basin, Italy (Pliocene and Quaternary tufo), ground water flows through openings generated by primary porosity in carbonate rocks. This type of ground-water circulation can be studied in the same way as that for rocks having interstitial porosity. In carbonate rocks as a whole the same zones are found as those in systems having interstitial porosity, but with some differences.

The zone of aeration (unsaturated zone, vadose zone). This is a zone through which rainwater or surface runoff infiltrates downward from the surface.

Infiltrating water runs over the surface of the limestone before it enters fissures (Figs. 14.3.3.1a and b) and follows a downward path until it reaches an impervious stratum or the water-table. To the above must be added water inflows by way of the shafts under dolines and water from stream losses. Within the vadose or unsaturated zone there are intermediary cavities which may temporarily hold the infiltrated water and dispatch it downwards in larger conduits. Thus in the zone of aeration the water can either move downwards rapidly as far as the main flow level by way of major openings in which losses are small or it can move very slowly through the fine cracks with a high retentive capacity.

Zone of underground stream flow. A zone of underground stream flow occurs where the discharge capacity of the permeable stratum is greater than the recharge rate. Thus ground water will not accumulate. Such cases may occur in the upper end of an aquifer system or where the system is perched, and in particular where it lies on an impervious stratum. The water flows on gravitationally by way of fissures, channels or galleries.

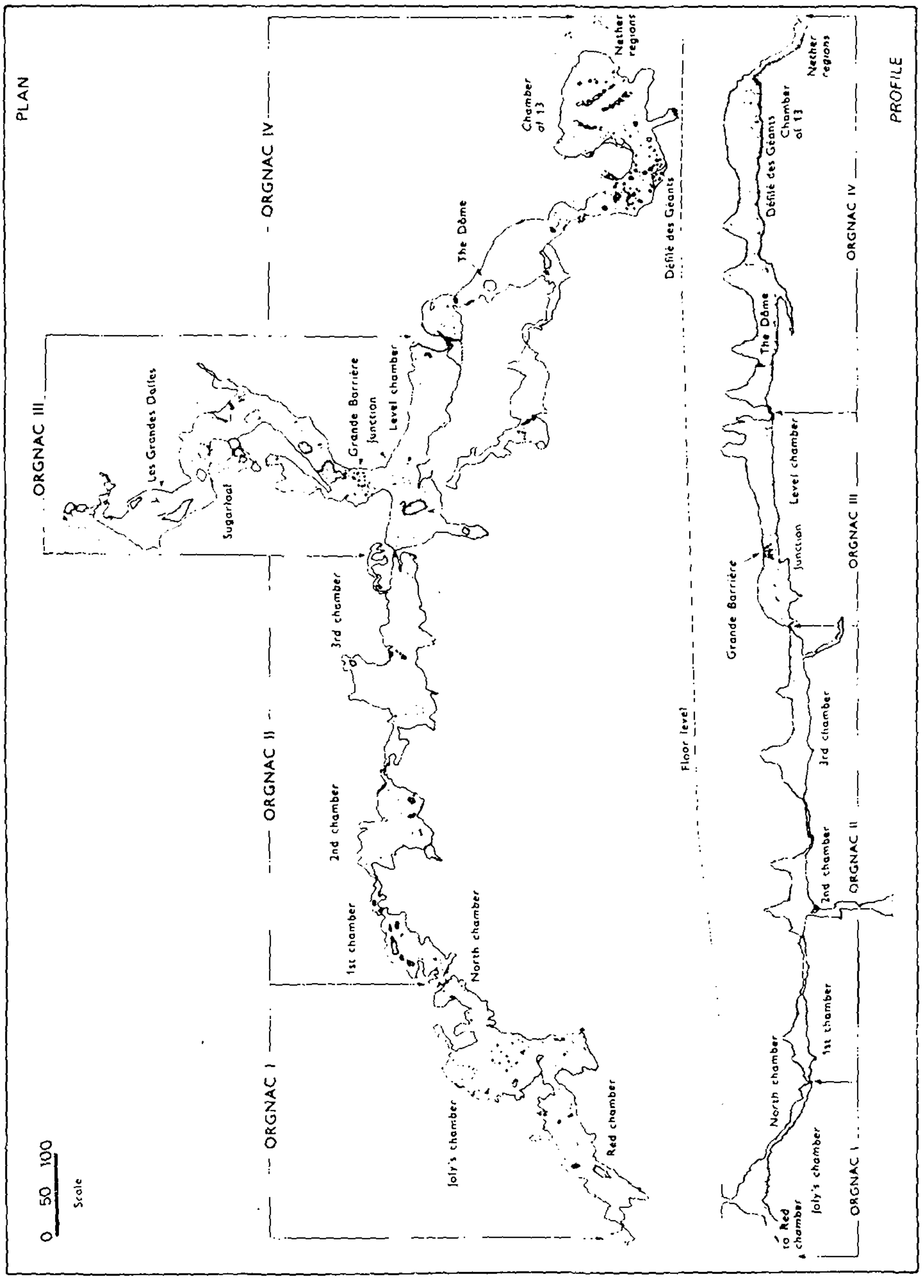

FIG. 14.3.2.2d. Orgnac Cave.

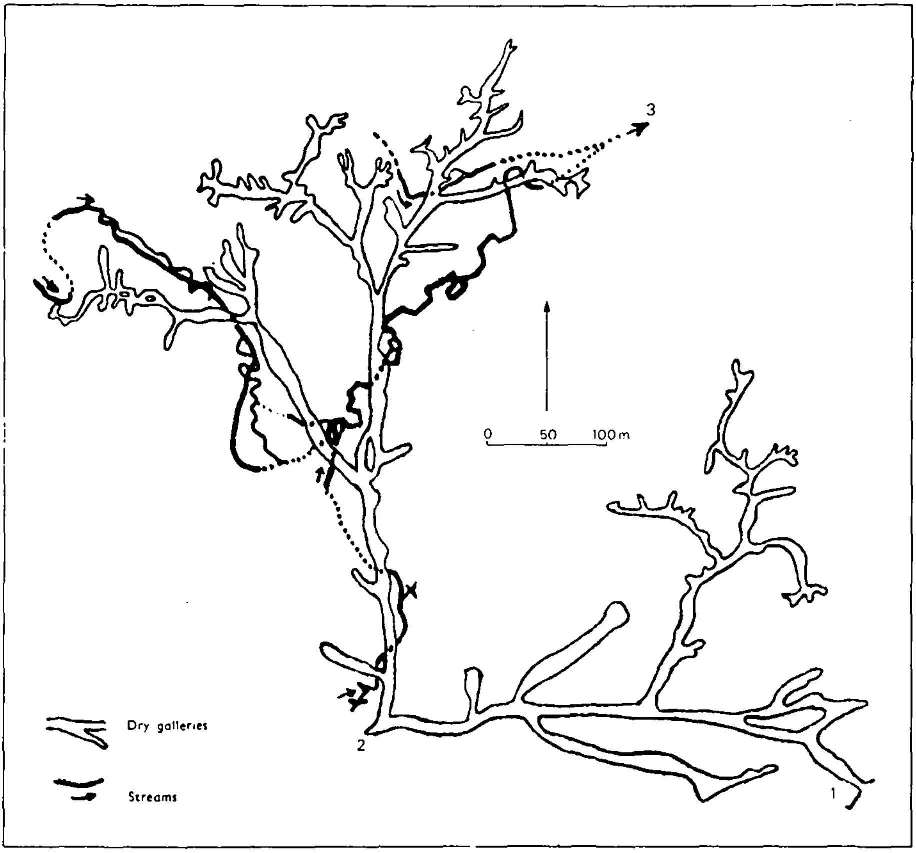

FIG. 14.3.2.2e. Miremont Cave (according to B. Pierret), showing superposed galleries of two successive karst systems: 1. Present entrance; 2. Possible outlet of former karst system; 3. Outlet of present karst system. (Cavaillé, 1964.)

FIG. 14.3.2.2f. Lattice labyrinth: Chaos Cave 4, Meurthe-et-Moselle (France), according to Louis and Lehmuller. (Renault, 1967).

Zone of saturation (zone of accumulation or phreatic zone). Water in the zone of saturation may be either unconfined or confined.

In an unconfined aquifer, the top of the aquifer is not in contact with the confining layer; there is an upper part including the zone of fluctuation of the water-table, where the widest openings, passages, chambers and halls are found, with floors having the same slight inclination as the water-table, and gravitational flows often concentrated in widely separated channels, commonly with high discharge; in the lower part, flow moves slowly through a larger number of diffuse, narrow fissures of channels. Nevertheless, favoured flowlines with heavy discharge in large channels may occur in places in the lower part,

In a confined aquifer the conditions for active circulation of water are not as good as in an unconfined aquifer, especially if the aquifer has only a low pressure gradient. Moreover, openings are more numerous in the outcrop areas of carbonate formations as a result of the dissolving action of the infiltrating water. The dissolving action

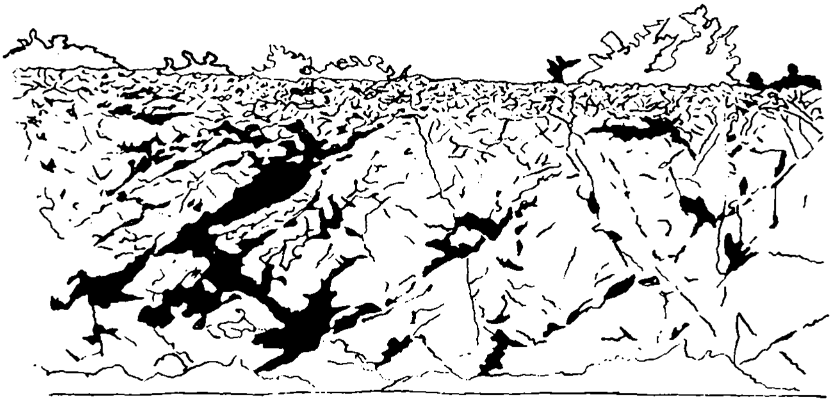

FIG. 14.3.3.1a. Eocene limestone in the cutting of a motor road near Zadar (Yugoslavia). (Vilimonović, 1965.)

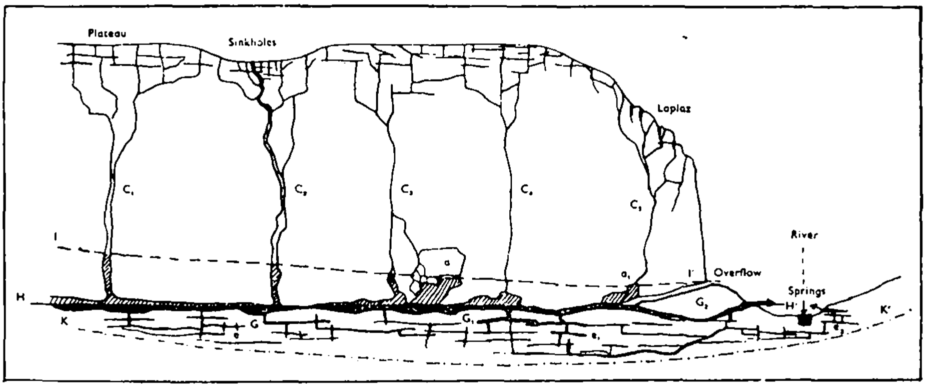

FIG. 14.3.3.1b. Diagram of karst system. Present flow of water in a homogeneous limestone mass. At the surface, fractures and infiltration joints; C–C_5, wells with rapidly descending water; G–G_2, horizontal galleries of karst flowage with chambers of rock detritus a; e–e_2, fractures and joints of the fissured network; K–K', level of karst basement; H–H', piezometric level of low water; I–I', maximum water level. (Cavaillé, 1964.)

diminishes rapidly down gradient. It is often observed that water circulation is active along lines of outcrop but is poor down dip and at depth.

14.3.3.2 Hydraulic conditions in carbonate rocks

To all appearances these laws are much more complex than those applying to terrains with interstitial porosity. The same laws do not apply for the vertical descent of infiltrating water, for gravity flow through channels and fissures, and for unconfined ground water and confined ground-water bodies.

As regards the vertical descent of infiltrating water, no general laws can be stated. Gravity flow through pores, fissures and channels depends on the width of the openings and the hydraulic head, the movement sometimes being oblique.

The laws of channel flow can be applied to gravity flow, in both the zones of aeration and of underground stream flow. This also applies to circulation in the upper part of the zone of saturation where the movement is concentrated in underground channels.

The average velocity (v) of the water and its rate of flow (Q), can then be expressed by Chezy's formula:

$$v = C\sqrt{R_h I}$$

whence $Q = C F\sqrt{R_h I}$, or $Q = C F^{1.5}\sqrt{I/P}$
in which the hydraulic radius $R_h = F/P$,
where:

F = cross-section of the channel;
P = wetted perimeter of the channel;
I = loss of head per unit of length $\Delta h/\Delta l$;
C = a coefficient depending on the nature of the channel sides.

Obviously the above formula can only be applied over short distances because of the rapid variations of F, P and also C. It can be assumed, however, that the flow is proportional to the square root of the loss of head. Sometimes the flow may become confined. In this case, it comes under the appropriate law for this type of channel (see the following sections).

A formula can be developed which can be applied to the flow of water in the saturated zone through fissures and through solution channels. Such flow may be either laminar or have a relatively high Reynolds number without becoming turbulent.

It has already been noted (see 14.1.4) that, in the case of a fissure with laminar flow, the specific discharge is proportional to the head loss:

$$q = K\, l\, I.$$

On the other hand, with flow having a high Reynolds number, the discharge is proportional to the square root of the head loss, as follows:

$$q = K' l \sqrt{I}.$$

In the case of a channel there are similar equations as follows: $q = K F I$ for laminar flow, and $q = K' F\sqrt{I}$ for flow with a high Reynolds number,
where:

K = hydraulic conductivity;
q = discharge;
l = width of the fissures;
F = cross-section of the channel;
I = $\Delta H/\Delta l$, the loss of head per unit of length.

In the saturation zone it is possible to have simultaneously laminar flow in the narrow openings and high Reynolds number flow at high velocity in the wide openings. These two types flow, taking place for one and the same head loss, are added together.

There then occurs in the narrow openings $q' = k'F'H/l'$, and in the large ones $q'' = k''F''\sqrt{H/l''}$ and $Q = q' + q'' = k'\,F'\,H/l' + k''\,F''\sqrt{H/l''}$.

In the above equations, k' and k'' relate to the cross-sections (F', F'') of the openings. It is possible, however, to choose the coefficients K' and K'' in relation to the cross-section F of the aquifer. In that case the total discharge is:

$$Q = K'F\frac{H}{l'} + K''F\sqrt{\frac{H}{l''}}.$$

If it is assumed that the lengths of trajectories are the same, then:

$$Q = K'F\,I + K''F\sqrt{I}.$$

Otherwise it would be:

$$Q = \frac{K'}{l'}FH + \frac{K''F}{l''}\sqrt{H} = K'_aFH + K''_aF\sqrt{H}.$$

For slight variations of Q and of I, $Q = K\,F\,I^{1/n}$.

An extensive carbonate aquifer can be treated as an aquifer with interstitial porosity, the only difference being one of scale. Accordingly, where intersecting fissures are closely spaced, the equation for an unconfined aquifer can be supplied (see Fig. 14.3.3.2a):

$$Q^n = K^n\frac{h^{(n+1)} - h^{(n+1)}}{(n+1)\,(x'-x)} = K^n\frac{H^{(n+1)} - h^{(n+1)}}{(n+1)\,X},$$

the value of n varying between 1 for laminar and 2 for turbulent flows.

For confined ground-water units (Fig. 14.3.3.2b) having a thickness m, a dip α, l being measured along the strata, and x in the horizontal line,

$$Q^n = K^n m^n\frac{H}{l} = K^n m^n\frac{H}{x}\cos\alpha.$$

The permeability in carbonate rocks is usually anisotropic. Permeability is far greater along the direction of the flow than transverse to it, because the flow enlarges the openings, in particular by dissolution.

14.3.3.3 Permeability and transmissivity

It follows from the above equations that both the concepts of coefficients of permeability (K) and coefficients of transmissivity (T) are acceptable in carbonate rock. These coefficients will be proportionately higher as karstification is more pronounced. In the absence of karstification, their values can be very low.

Some examples of permeability are given below (K(cm/s)):

Tertiary limestone from South Carolina (Siple, 1967): near Savannah, 2.24×10^{-2}; region of Burton, 0.56×10^{-2} to 5.66×10^{-2}; region west of Broad River, 4.5×10^{-2} to 10.2×10^{-2}; fifteen miles east of Savannah, 5.5×10^{-2}.

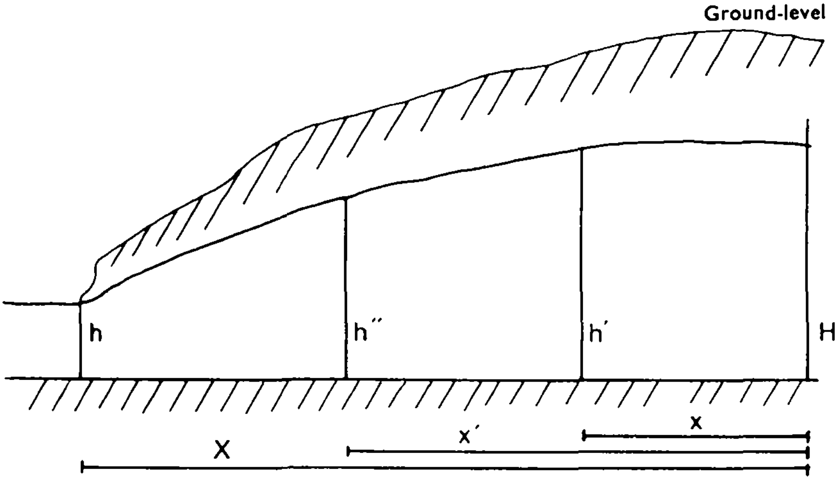

FIG. 14.3.3.2a. Unconfined aquifer.

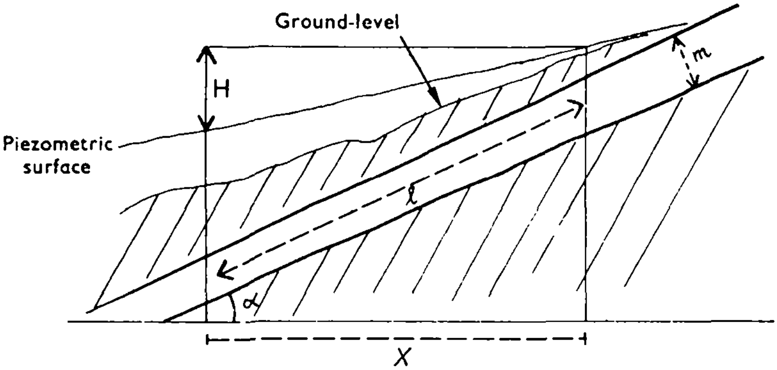

FIG. 14.3.3.2b. Confined aquifer.

Tunisia (Gosselin and Schoeller, 1939): Campanian (Upper Cretaceous), Ketena well, 88.4 × 10^{-2}; Mareth well, 11.6 × 10^{-2}; Cedria well, 6.5 × 10^{-2}.

Eocene, Aïn Faouar well, 1.23 × 10^{-2}; Aïn Tounga well, 1.01 × 10^{-2}; Aïn Cherfi well, 0.06 × 10^{-2}.

For a confined aquifer, the storage coefficients are generally between 1 × 10^{-2} and 5 × 10^{-4}. For unconfined water the values will be much greater, even as high as 0.25.

Aquifer porosity. Important here is not the interstitial porosity of the porous limestones, but the aggregate volume of open spaces in cracks, fissures, channels, galleries, etc.

This type of porosity varies enormously, both laterally and vertically, within the same hydrogeological conditions. It may be high in the zone of aeration, and also in the zone of water-table fluctuation, decreasing towards the bottom of the zone of saturation.

The high porosity near the intake area of a confined aquifer decreases down-gradient where, in certain cases, it can be extremely low. Conversely, it can be high down-gradient if a palaeokarst is present.

Porosity is difficult to measure. In the Silurian-Devonian of south-east Spain (Navarro and Samper, 1965) it ranges from 10 to 12 per cent; in the Middle Trias dolomite from 6 to 7 per cent. There are instances, however, of near-zero porosity and others where it may be as high as 30 per cent.

The frequency of channels is greater in the upper part of the zone of saturation than in the lower part. At the Fontaine de Vaucluse in south-eastern France the porosity at the top of the zone has been estimated at 13 to 14 per cent (Schoeller and Aigrot, 1967) and that of the part immediately below at 0.2 to 0.8 per cent. These estimates were made by measuring the variations in discharge. In 1969, Flandrin and Paloc obtained a porosity value of 1.2 per cent by studying the relationship between decline in level from an observation well and the decrease in discharge from the spring.

Study of the discharges of springs has shown that the volume of the narrow openings far exceeds that of the wide openings. Thus (Schoeller, 1967), the proportions in the respective supplying massifs are as follows:
at the Fontaine de Vaucluse, 0.31 wide openings compared to 0.69 narrow openings;
at the Spring of Kef (Tunisia), 0.20–0.25 compared to 0.80–0.75;
and at the Spring of Lez Hérault, 0.24 compared to 0.76.

14.3.3.4 Geographical and geological aspects

Water circulation in carbonate rocks. The numerous hydrological behaviour patterns in carbonate rocks result from the diversity of ambient geographical and geological conditions. Water circulation also depends on the physical and chemical character of carbonate rocks which may be: (a) compact, thick limestones, with circulation essentially taking place in widely separated, large openings or main channels; (b) bedded limestones, with circulation either in narrow or enlarged fissures and hence with a possibility for diffuse circulation and of drainage by large conduits; (c) relatively thin bedded limestones or limestones alternating with marly beds to form poor aquifers; (d) chalk formations, with great interstitial porosity, but which may be impermeable in the mass—in which case the water will flow only in conduits; (e) dolomite.

The hydrology of carbonate rocks and that of what is called karst are intimately interconnected. The term karst unites specific morphological and hydrological features in soluble (mostly carbonate) rocks. Morphological features include 'basins of closed drainage, lost rivers, estavellas, vauclusian springs, more or less individual underground streams and incongruity of surface and underground divides'.[1] But the meaning of karst has been enlarged to define 'a carbonate rock whose openings have been enlarged by the action of ground water' (Burdon and Papakis, 1963).

Cvijic (1960) made the following distinction: (a) 'complete' karsts in which all the karstic forms are perfectly developed, as are the most varied karst phenomena. Known as holokarst, it develops in the pure, thick limestone extending from the top of the formation down to great depth—even below the lowest topographical level; (b) Causse karst,[2] in which a great limestone mass rests upon an impervious base outcropping in at least one valley; (c) an imperfect karst, known as merokarst, in which only certain karst characteristics have been developed. It retains considerable surface drainage, and is formed from non-homogeneous strata, including marly limestones and dolomites; (d) a Jura karst[3] formed from alternating thick limestone beds and marly strata cut by deep valleys, irrespective of whether the terrain is folded or not.

There is another aspect, however, which is not to be neglected for it conditions the ground-water storage:

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[1] *Glossary and Multilingual Equivalents of Karst Terms*, 1st. ed., Rome, FAO, 1971.

[2] The Causses is a karstic region in the southern part of France.

[3] The Jura is a mountaineous region of France and Switzerland.

Regions with outcropping carbonate rocks.

1. Bare carbonate rocks provide rapid infiltration and low evaporation losses. The infiltrating water contains little CO_2, and consequently its aggressivity is reduced.

2. Carbonate rocks with a cover of soil and vegetation. There is soil cover, however thin. Infiltration is retarded. Moreover, there is water loss by evapotranspiration, which is the greater in proportion as the soil cover is thicker and the plant cover denser. The water picks up CO_2 in the soil and is highly aggressive.

Buried carbonate rocks. In an unconfined aquifer:

1. If the cover is permeable, formed for example by unconsolidated sand, alluvia, etc., the water from the cover may be drained by the limestone. A recharge reserve would be thus created, ensuring a certain regularity of flow.

2. If the cover is impermeable, water could not readily move into underlying limestone. Consequently, water intake and karstification would be impeded.

Carbonate rocks in extensive geological units. To evaluate water resources and how to exploit them, the diverse karst conditions in major geological units need to be examined (Maksimovic, 1969).

1. Shield karsts are for the most part completely eroded. Metamorphosed limestone involved in ancient folding processes is accordingly of less importance than that in platforms and mountain ranges. Differences in altitude between intake zones and outflow zones are small. The karst limestones are of limited areal extent and are not highly karstified.

2. Platform karst has individual characteristics differing appreciably from those of the karsts in shields or in folded mountain chains. It can be either an open karst or a closed karst.

(a) *Open karst* springs may flow at the surface, or into material buried under a thin cover. Circulation is active.

Plateau karst has deep valleys cut down to impervious strata, and all circulation occurs above stream or sea level. The limestone aquifer rests upon an impervious stratum. Usually in a karst of this type there is only a zone of underground stream flow, and rarely a zone of saturation. Water circulates along almost horizontal galleries with upgradient branches. The galleries may continue for considerable distances without any appreciable change in width. Each of the widely separated passages ends with a very constricted exit though which the large flow is generally above the level of the valley floor. In the Padirac Cave (France) there is a single gallery, and in that of Miremont (France), a dendriform gallery branching upgradient (Fig. 14.3.2.2e).

(b) *Closed ground-water systems.* 'Plains karst' differs from plateau karst in that there is circulation down-gradient below valley base level. Plains karst occurs in flat country or in regions where the valleys are not eroded down to an impervious base. Springs in the valleys are artesian, and the lower part of the aquifer forms a ground-water reserve. The nature and density of the openings depend on the past history of the particular karst. If it was originally a plateau karst with valleys which developed for a long time and were later filled in, there will be only widely spread openings. If the karst has always been below the valley floor, the openings may be more dispersed in a down stream direction. Ground water in this type of karst may circulate beyond the margins of the valley, passing beneath the interfluves.

(c) *Carbonate rock artesian systems.* The platforms incorporate vast sedimentary basins with artesian or confined ground-water bodies. The pressure water basins have an intake region consisting of:

(i) An external zone upstream from the carbonate rock aquifer, such as an impervious clayey or metamorphic formation. The surface water from this zone can be absorbed by infiltration or by disappearing into caverns at the edge of the karst, into which rivers can vanish. In cuestas, where the base of the limestone is above the level of the external zone, the external intake zone does not extend around the entire periphery. Losses of water will occur only where the base of the limestone happens to be cut by a valley.

(ii) An area of recharge at the head of the confined aquifer, where the limestone outcrops occur. In the whole of this area, marked karstification will be observed, comparable to that of the plateau and plains karsts.

(iii) An area in the upper part of the confined zone, where there may also be some karstification caused by the residual aggressivity of the recharge water.

(iv) The major confined zone extending down gradient, where karstification may be almost non-existent, unless a paleokarst existed previously. The flow is diffuse, slow and often limited.

The narrow fissured conduits and low potentiometric gradient tend to restrict circulation, so that the absorption capacity of the confined aquifer is very much less than the absolute infiltration capacity.[1] As a consequence, springs develop at the limit of the intake areas of the confined aquifer, and the ground water tends to circulate parallel to the outcrops with water moving in the lower section of the intake area and in the upper part of the confined zone.

3. Ground-water systems in folded regions. In the case of burried synclines where an impervious stratum covers the limestone, the limestone water is confined and circulation is difficult, which reduces solution. Only the outcrops and confined rocks near by can be easily karstified. Overflow springs may occur, the absorption capacity being lower than the potential infiltration capacity. Furthermore, if the flank outcrops have a plunge parallel to the axis of the fold, a flow will tend to develop along the same direction.

As in the case of a holokarst, a succession of anticlines and synclines may develop parallel to each other, forming a relatively complex tectonic system but a relatively homogeneous hydrogeological one. The underground circulation may be parallel to the axis of the folds, and sometimes a synclinally folded limestones overlying an impermeable outcropping bed will be easily karstified because ground water circulates easily. If the syncline plunges, there may be only one spring at the lower end. If one of the synclinal flanks is markedly lower than the other, there may be direct movement of water from one flank to the other; or if only one flank has been preserved, the karstification will form descending passages almost parallel to the dip.

In mountain massifs, blocks of limestone may in places be enclosed by crystalline or metamorphic blocks. On a smaller scale, isolated blocks may rest on an impervious basement. In this case, karstification may resemble that of the plateaux.

In mountain karst (Fig. 14.3.2.2b), the zone of aeration forms the major part of the aquifer. The potentiometric surfaces adjust to the levels of the springs. In most cases there is a great difference in altitude between the levels of the springs and the intakes. Many are hillside springs, a comparable number occur in the valley bottoms. A mountain karst may therefore have no well-developed zone of saturation.

The passages have a very pronounced slope, approximating to that of the natural surface, and are not filled by alluvium, because they are flushed out by torrential flow. Additionally, they may carry the water from the intake area directly to the springs. In the zone of aeration, there are often large chambers and the complete network is labyrinthiform, as a result of the reductions of stress which ablation produces in the limestone mass.

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[1] The absorption capacity of an aquifer is the maximum amount the ground-water body is able to carry from the recharge to the discharge area. It is related only to the transmissivity of the aquifer and the potentiometric gradient. The absolute infiltration capacity is the maximum amount of rainwater a recharge area is able to infiltrate. H. Schoeller, *Reading Symposium. World Water Balance*, p. 60 International Association of Scientific Hydrology, 1970.

14.3.3.5 Springs

Generally speaking, carbonate rock springs represent constricted discharges, with abundant flows at widely separated outlets. However, there is no emergence of the water-table, in the sense of a spring extending uninterruptedly along a wide stretch of aquifer front and with a feeble discharge per unit of front. This is especially so in a strongly developed karstic form (Mandel, 1967). At the start, lines of weakly discharging springs may develop. The largest spring progressively captures the water feeding its neighbours, and they become dry.

A resurgence is the reappearance of a stream which crosses an intake and disappears underground, to reappear at the surface some distance downstream. An exsurgence is a spring fed entirely by seepage.

Springs are classified as follows (Fig. 14.3.3.5a):

Unconfined ground-water springs

Contact springs. 'A contact spring is one whose water flows to the surface from permeable material, over the outcrop of less permeable or impermeable material that retards or prevents the downward percolation and thus deflects it to the surface' (Meinzer, 1923a).

Simple flowing springs. All the flow lines upstream from the spring are above an impermeable layer.

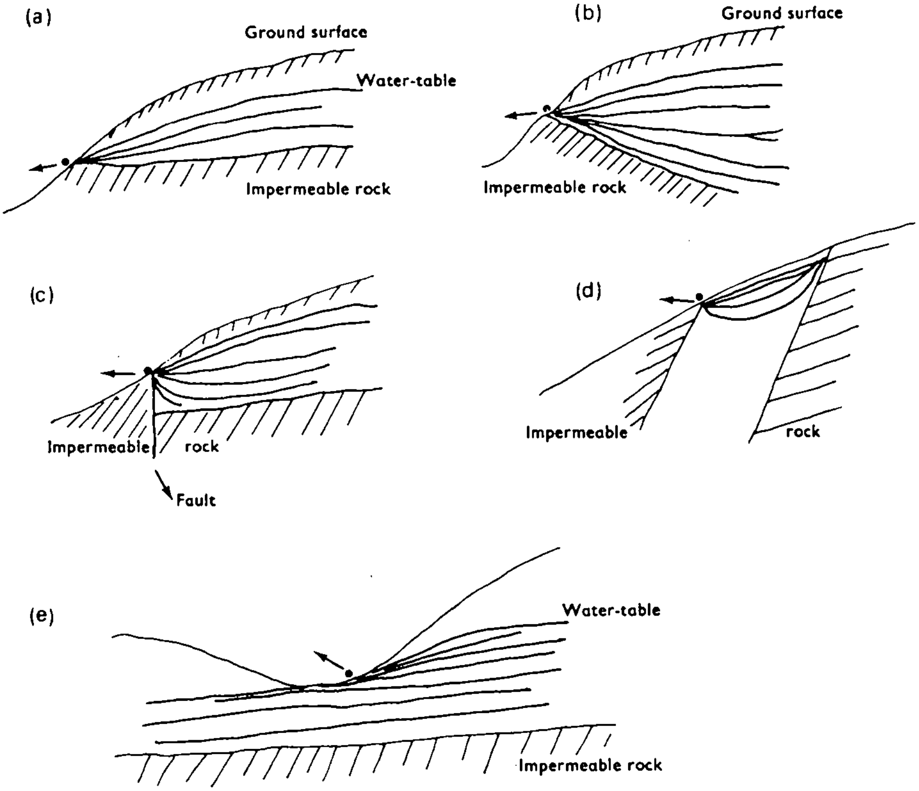

FIG. 14.3.3.5a. Springs: (a) simple flowing spring; (b), (c) overtopped springs; (d) overfill spring; (e) depression spring.

Overtopped springs. Near the spring, the outflow lines are forced to the surface.

Overfill springs. These are overflowing springs at the top of a confined ground-water body.

Depression springs. 'A depression spring is one whose water flows to the surface from permeable material, simply because the surface extends down to the water table' (Meinzer, 1923a).

Confined ground-water springs

Artesian springs. 'An artesian spring is one whose water issues under artesian pressure, generally through some fissure or other opening in the confined bed that overlies the aquifer. Such springs, by definition, occur only in areas of artesian flow, where the piezometric surface is above the land surface' (Meinzer, 1923a).

The presence and state of a spring are often determined by fractures, faults or tectonic cracks, and also by joints. On the side of a valley, the karst water escapes through a labyrinthine complex of fissures parallel to the valley flanks. Underneath the valley, pressure-relief joints parallel to the valley floor enable the water to emerge some distance from the toe of the slope (Renault, 1967, 1968, 1970).

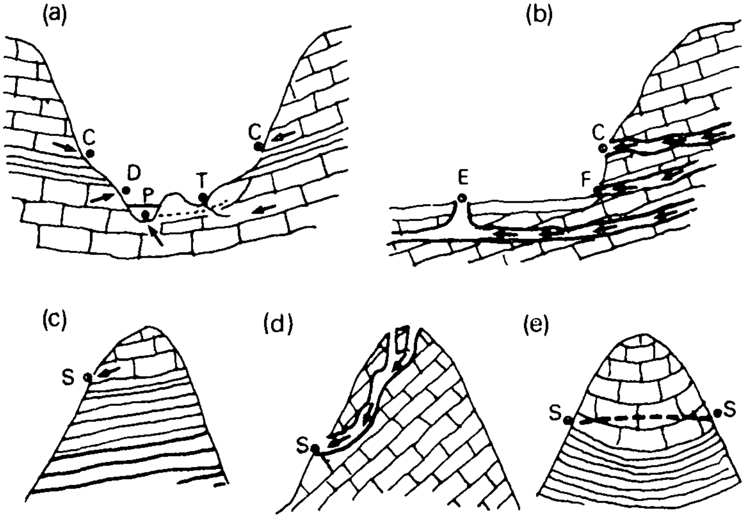

Fig. 14.3.3.5b. Karstic springs of denudation: (a) Erosion spring; C, on a slope; D, in doline; T, on a terrace; P, in a river. (b) Polje spring; C, on a slope; F, at the foot of a mountain; E, estavelle. (c) Summit spring. (d) Slope spring. (e) Overflow spring. (Maksimovitch, 1966.)

Several types of springs (Maksimovic, 1966) can be distinguished, as follows (Figs. 14.3.3.5b–d):

Erosion springs (produced by stream flow) consist of slope-erosion springs (on plains, plateaux, mountains); erosion springs occurring above an impervious stratum; erosion springs in valley bottoms, on terraces, and in river beds.

Denudation springs (caused by surface runoff) are found on slopes and on crests.

Polje springs consist of common springs and estavelles.

Barrier springs are caused by an impervious barrier, down gradient, and include fault springs and coastal springs.

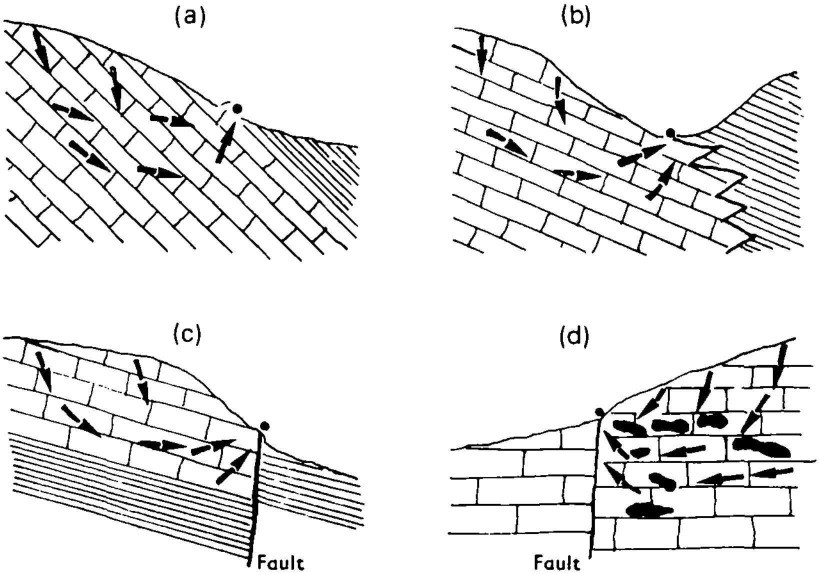

FIG. 14.3.3.5c. Karst springs of dams or faults. Lithological dam formed by non-karstifiable rocks: (a) covering; (b) of changing facies. Fault-dam spring: (c) formed by non-karstifiable rocks; (d) formed by karstifiable though non-karstified rocks. (Maksimovitch, 1966.)

14.3.4 *Ground-water chemistry*

The real characteristic of carbonate rock hydrogeology is the presence of numerous and perhaps wide cavities through which water circulates or has circulated. The cavities are genetically developed by water circulation, and are formed by the attack of dissolved carbonic acid on the limestone.

14.3.4.1 **Dissolution of limestone**

Only a minute proportion of the carbonic acid in solution comes from the atmosphere, the remainder from the air contained in the soil. Indeed, the CO_2 pressure in atmospheric air is only 0.0003 atmosphere, against 0.001 to 0.1 atmosphere in soil air.

The process of dissolving limestone is represented as follows:

1. Dissolving of CO_2

$$[CO_2] = \frac{\alpha_\mu \delta}{22.4} = \frac{\alpha_\mu C P}{22.4} \qquad 14.3(1)$$

where:

$[CO_2]$ = the concentration in moles of CO_2 in the water;
α_μ = the coefficient of absorption for a given temperature and a given ionic strength μ;

δ = the partial pressure of the CO_2 in the atmosphere;
C = concentration of CO_2 in the atmosphere;
P = the pressure in atmospheres of the sum of the gases.

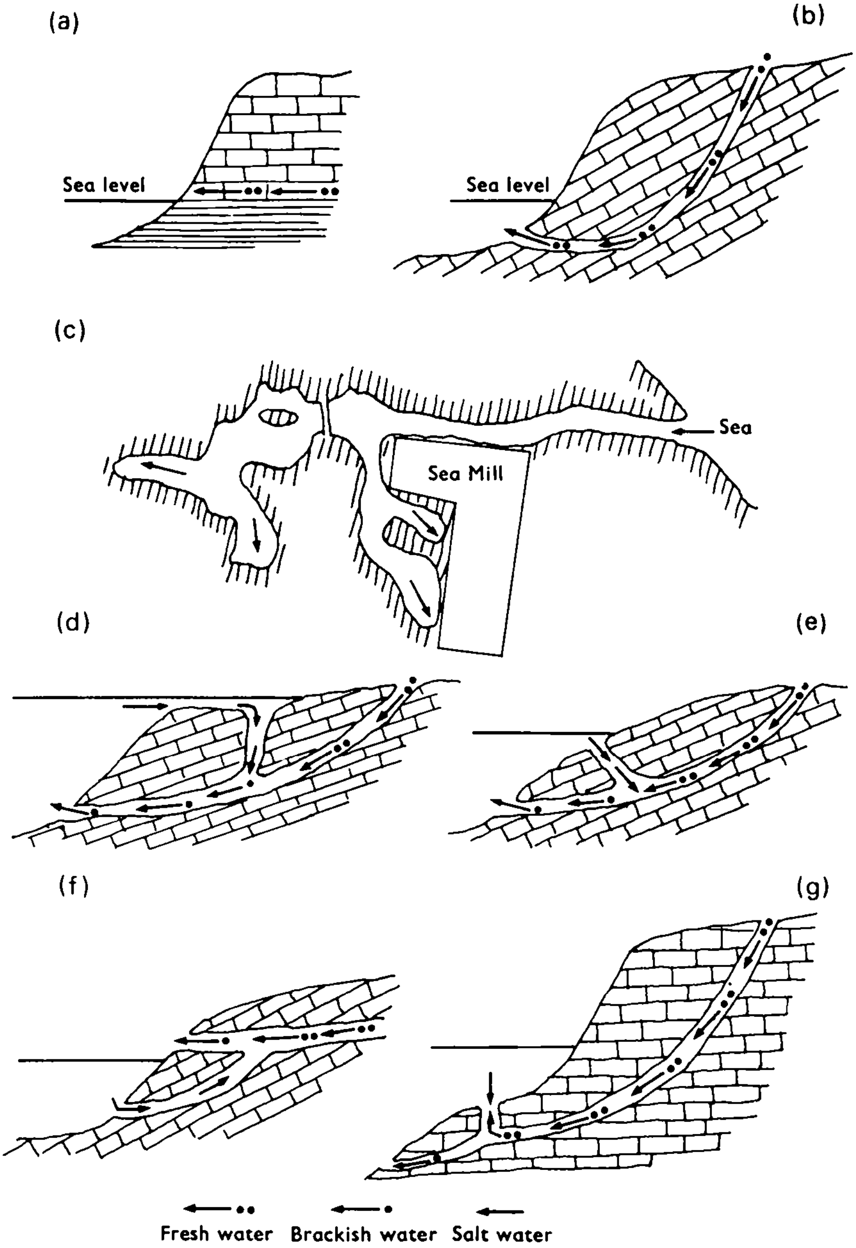

FIG. 14.3.3.5d. Coastal springs: (a)abrasion spring; (b) submarine fresh-water springs; (c) sea-mills at Kephalina, according to Martel, 1908; (d) diagram of a sea-mill; (e) sea chasm; (f) brackish water spring; (g) submarine spring (intermittent). (Maksimovitch, 1966.)

2. Hydration balance:

$$CO_2 + H_2O \rightleftarrows H_2CO_3.$$

3. Dissociation of carbonic acid balance:

$$H_2CO_3 \rightleftarrows H^+ + HCO_3^-.$$

4. Dissociation of ion bicarbonate balance:

$$HCO_3^- \rightleftarrows H^+ + CO_3^=.$$

5. Dissociation of $CaCO_3$:

$$CaCO_3 \rightleftarrows Ca^{++} + CO_3^=.$$

6. Dissociation of water:

$$H_2O \rightleftarrows H^+ + OH^-.$$

This leads to the following equations:

$$\frac{\langle H^+\rangle [HCO_3^-]}{[CO_2]} = \frac{\gamma_0}{\gamma_1} K_1 = K_1'''. \qquad 14.3(2)$$

Here CO_2 represents the dissolved CO_2 plus the acid H_2CO_3.

$$\frac{\langle H^+\rangle [CO_3^=]}{[HCO_3^-]} = \frac{\gamma_1}{\gamma_2} K_2 = K_2''' \qquad 14.3(3)$$

$$[CO_3^=]\ [Ca^{++}] = \frac{K_c}{\gamma_2\ \gamma_{Ca}} = K_c' \qquad 14.3(4)$$

$$\langle H^+\rangle\ \langle OH^-\rangle = K_{H_2O} \qquad 14.3(5)$$

$$pH = \log \frac{1}{\langle H^+\rangle} \qquad 14.3(6)$$

The pointed brackets $\langle\ \rangle$ represent activity concentrations; the square brackets [ ] molal concentrations; K is a constant of dissociation; γ_0, γ_1, γ_2, γ_{Ca} are the activity factors of CO_2, HCO_3^-, $CO_3^=$, and Ca^{++}.

By combining the above equations one obtains the CO_2, δ, and pH equilibrium.

$$[CO_2]_{eq} = \frac{K_2'''}{K_1'''\ K_c'} [HCO_3]^2\ [Ca^{++}] \qquad 14.3(7)$$

$$\delta_{eq} = C.P._{eq} = \frac{22.4}{\alpha_\mu}\ \frac{K_2'''}{K_1'''\ K_c'} [HCO_3^-]^2\ [Ca^{++}] \qquad 14.3(8)$$

$$[CO_2]_{eq} = \frac{K'_c}{K'''_1 K'''_2} \; \frac{(H^+)^2}{[Ca^{++}]} \qquad 14.3(9)$$

$$pH_{eq} = \log \frac{K'_c}{K'''_2} - \log [HCO_3^-] - \log [Ca^{++}] \qquad 14.3(10)$$

$$pH_{act} = \log \frac{1}{K'''_1} + \log [HCO_3^-] - \log [CO_2] \qquad 14.3(11)$$

where pH_{act} is the actual pH measured by electrometric or by colour method.

At a temperature of 15° C the following values are obtained for $\log K'''_1$ and $\log K'_c/K'''_2$:

μ	0.001	0.0036	0.01	0.04	0.1
Log K'''_1	$\bar{7}.60$	$\bar{7}.61$	$\bar{7}.63$	$\bar{7}.63$	$\bar{7}.70$
Log K_c/K'''_2	2.13	2.19	2.28	2.44	2.58

and for μ = 0.001 at different temperatures:

t	10° C	15° C	20° C	25° C	30° C
log K'''_1	$\bar{7}.55$	$\bar{7}.60$	$\bar{7}.63$	$\bar{7}.67$	$\bar{7}.69$
log K_c/K'''_2	2.21	2.13	2.08	2.00	1.92

μ is the ionic strength:

$$\mu = \tfrac{1}{2} [C_1 \times Z_1^2 + C_2 \times Z_2^2 + \dots + C_n \times Z_n^2],$$

C being the molal concentration and Z the valence of each ion.
μ can be calculated by means of milliequivalents r:

$$\mu = [rCa + rMg + rFe + SO_4 + rCO_3 + \frac{(rNa + rK + rCl + rNO_3 + rHCO_3)}{2} \times 10^{-3}$$

and for $\log K'''_2/K'''_1 K'_c$ the following is obtained with:

μ	$\log K_2'''/K_1'''K'_c$
0	$4.30 + 0.0092\ t$
0.001	$4.21 + 0.01517\ t$
0.0036	$4.14 + 0.01504\ t$
0.01	$4.05 + 0.01487\ t$
0.04	$3.87 + 0.01424\ t$
0.1	$3.68 + 0.01423\ t$

in which t is the temperature in Celsius degrees between 0° and 50°.

14.3.4.2 Dissolution of dolomite

The dolomite solubility is ruled by the solubility product:

$$[Ca^{++}]\,[Mg^{++}]\,[CO_3^{=}]^2 = \frac{K_d}{\gamma_{Ca}\;\gamma_{Mg}\;\gamma_2^2} \qquad 14.3(12)$$

which gives, taking into account equation 14.3(3):

$$[Ca^{++}]\,[Mg^{++}] = \frac{K_d}{K'_2}\;\frac{\langle H^+\rangle}{\gamma_{Ca}\;\gamma_{Mg}\;\gamma_1^2\,[HCO_3^-]^2}. \qquad 14.3(13)$$

When the equilibrium is reached between the liquid and solid phase of a rock containing both $CaCO_3$ and (Ca, Mg)CO_3, equations 14.3(4) and 14.3(12) give:

$$\frac{[Mg^{++}]}{[Ca^{++}]} = \frac{\gamma_{Ca}}{\gamma_{Mg}}\;\frac{K_d}{K_c^2} \quad \text{or} \quad \frac{\langle Mg^{++}\rangle}{\langle Ca^{++}\rangle} = \frac{K_d}{K_c^2}. \qquad 14.3(14)$$

Then, according to the values of K_d and K_c given by Langmuir (1971), we have, for water flowing through dolomite $Ca_{0.50}Mg_{0.50}CO_3$, the ratios given in column I of the table below. But there is a more soluble dolomite $CaO_{0.54}MgO_{0.46}CO_3$ (Langmuir, 1971) with $K_d = 3.24 \times 10^{-17}$ at 10° C. The ratios for this dolomite are given in column II.

Temperature (°C)	I $<Mg^{++}>/<Ca^{++}>$	II $<Mg^{++}>/<Ca^{++}>$
0	1.32	2.78
5	1.12	2.13
10	1.01	1.67
15	0.89	1.32
20	0.76	1.08
25	0.63	0.89

The ratios $[Mg^{++}]/[Ca^{++}]$ may be 2 or 3 per cent greater than those of $\langle Mg^{++}\rangle/\langle Ca^{++}\rangle$ because of $\gamma_{Ca}\gamma_{Mg}$ and ion pairing of Ca and Mg in the forms of $MgSO_4^{\circ}$, $MgHCO_3$, $MgCO_3^{\circ}$.

Ground water coming from limestone has a ratio [Mg]/[Ca] or rMg/rCa lower than the values of the preceding table, minus the 2 or 3 per cent mentioned above. When $[Mg^{++}]/[Ca^{++}]$ is greater, the rock contains Mg, in the form of ($MgSO_4$, $MgCO_3$) salts, except when there has been an Mg increment through base exchange.

It is observed that, in water from the same kinds of terrains, the Mg/Ca ratio is fairly constant.

14.3.4.3 Aggressivity of water

Water is aggressive when its dissolved CO_2 content is greater than the equilibrium CO_2 content given by equations 14.3(7) and 14.3(9), or when the CO_2 partial pressure exceeds that given by equation 14.3(8), and also when the actual pH is less than the equilibrium pH of equation 14.3(10).

14.3.4.4 Precipitation of $CaCO_3$

Precipitation of $CaCO_3$ begins under the opposite conditions, when CO_2 is released from solution until it reaches a value below equilibrium. The different equilibrium equations show that precipitation may occur through a Ca increment caused by base exchange, and by the dissolving of $CaSO_4$.

14.3.4.5 The effect of chemical phenomena in channel formation

The above chemical formulae permit a better understanding of how cavities are formed in carbonate rocks.

First, there must be pre-existing fissures because no enlargement can occur without the movement of water. Furthermore, the fissures must be reasonably wide. A fissure less than 0.25 mm in width will not become lengthened or enlarged because the circulating water rapidly becomes saturated with $CaCO_3$. A fracture or a capillary tube of 1 mm width or radius will allow water to flow along its length to a depth of several hundred metres (Weyl, 1957).

Water dissolves CO_2 from the overlying soil, producing carbonic acid, which provides H^+ ions. The organic acids in the soil also contribute H^+ ions, so that the water becomes highly aggressive. Below soil level, this aggressive water attacks the limestone surface and penetrates into the fissures. Some aggressive water may also reach the fissures directly from the soil. This process reduces the aggressivity of the water, but enough remains to permit enlargement of the fissures in the zone of aeration. Even when water is not percolating downwards, the CO_2 gas will itself penetrate the fissures, and be dissolved in the ground water, thus maintaining by reason of its greater density and by its dissolution a degree of aggressivity.[1]

When the descending water reaches the roofs of cavities, it comes into contact with air. The escape of gaseous CO_2 then leads to supersaturation and the precipitation of $CaCO_3$, especially when the cavities are well ventilated.

When the water moves down into the zone of saturation, it can do so either by direct filtration, or by moving through open cavities. In the zone of saturation, the ground water is nearly always aggressive. This aggressiveness may be residual, or may be caused by the dissolving of CO_2 moving downwards from the soil above.

Whenever the barometric pressure falls in the Lascaux Cave (France), air charged with CO_2 is released under pressure from where it is entrapped in a sealed cavern below. It is also arguable that CO_2 may come from deeper waters at higher temperatures which reduce the coefficient α of CO_2 absorption. This suggests that precipitation of $CaCO_3$ may occur at greater depths.

14.3.4.6 General water quality

Generally speaking, the unconfined ground-water bodies in carbonate rocks are chemically low in dissolved solids. This is because limestone dissolves only in a limited way, conditioned by the CO_2 partial pressure of the soil air, which itself varies only within narrow limits. For this reason the $CaCO_3$ content rarely exceeds 350 ppm. If there is only pure limestone, other dissolved salts will not be present in large quantities because these can only be dissolved at the same time as the actual limestone. For this reason most of the non-carbonate salts such as chlorides and sulphates are derived from rainwater concentrated by evapotranspiration. Denuded carbonate rocks commonly contain water having a low salinity.

This situation is not true for bacteriological qualities, because filtration is very poor, which makes limestones particularly susceptible to pollution.

It is imperative that regular pollution checks be made on waters occurring below bare karst surfaces, especially where there are dolines, shafts, sinkholes, or downward leakage of surface water.

The chemical quality of confined water may also be excellent, particularly at the intake area. Even at some distance from the intake, it may continue to be good, when the water circulates freely either because the system is an open one or because it includes

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[1] As Weyl, Roques and other authors have shown, the large openings can only have been formed by large flows of water.

a paleokarst. Where the system is closed, however, or where circulation is difficult or even non-existent, salts such as chlorides and sulphates may become highly concentrated by solution from the limestones, even at great depth.

14.3.5 *Water resources*

The water resources of carbonate rocks depend not only on the volume of water that can be stored, but also on recharge, i.e. the quantity of water which can be replenished in a given time.

14.3.5.1 Recharge or replenishment

Recharge may be indirect or direct (14.3.3.2). Surface runoff is limited and may not occur. The existence of surface depressions and dolines is an indication that surface runoff may play no part in disposing of rainwater.

Condensation of the moisture in the air penetrating into fissures in the zone of aeration may contribute a certain amount of water. This will occur only in hot weather when even relatively dry air may be charged with moisture. When it penetrates into the cooler limestone, condensation can take place.

14.3.5.2 Recharge rate: recharge index

The proportion of rainfall recharging the ground water is variable, depending on whether the rock surface is bare, or has a plant cover. It also depends on the rainfall intensity, because heavy rain reduces the effect of evapotranspiration; and on the season, because evapotranspiration is greater in the dry season. However, in a prolonged drought the soil moisture becomes exhausted and the total evapotranspiration losses are less. This means that recharge rates or indexes vary considerably. The average and maximum rates and indexes are higher for a karst terrain than for any other. Examples are shown in Tables 14.3.5.1a and b.

14.3.5.3 Resources

These include natural resources which are replenished year by year, and those in storage which may be depleted by overpumping.

Natural resources. Natural resources can be calculated by the methods applicable to rocks with interstitial porosity (see Chapters 5 and 6).

However, for carbonate rocks the chloride method (Schoeller, 1962) is particularly applicable because surface runoff is practically nil, and the rock contributes only a very small quantity of Cl to the water. Almost the whole of the Cl comes from the rainwater. The Cl of the rainwater is concentrated in the soil by evapotranspiration, is carried down to the water-table and has much the same concentration when it emerges from springs, so that:

$$\frac{I}{P} = \frac{\mathrm{Cl}_p}{\mathrm{Cl}_n},$$

where:

I = the infiltration;
P = the rainfall;
Cl_p = the chloride concentration in the rainwater;
Cl_n = the chloride concentration in the ground water.

TABLE 14.3.5.1a. Recharge rates in various Mediterranean countries[1]

Country	Pmm	$Q_w$mm	$A = \frac{Q_w}{P} \cdot 100$
Tunisia (long-term period, see below)			
Djebel Chemnate	633	264	42
Ben Saidane	472	108	23
Zaghouan	463	166	36
Greece			
Lilaia group of springs	1,400	723	51.6
Parnassos-Ghiona region	1,150	583	50.7
Israel			
Western Galilee	750	229	30.5
Yargon basin	750	335	44.7
Na'man spring	600	318	53.0
Tunisia (example of annual variation)			
Le Kef 1927	463.5	154	33.2
1928	643.0	356	55.4
1929	616.0	554	90.0
1930	633.6	444	70.1
1931	513.7	415	80.8
1932	676.0	537	79.5
1933	348.2	266	76.4
1934	606.0	445	73.6

1. P is rainfall, Q_w is the rate of discharge of the spring, $Q_w$mm the discharge divided by the intake area.

TABLE 14.3.5.1b. Monthly variation in recharge rates in Hungary, averages taken over a period of twenty-two years (Kessler, 1967)

Month	A (percentage)	Month	A (percentage)
January	50.2	July	20.6
February	73.3	August	18.1
March	123.6[1]	September	15.6
April	65.4	October	12.3
May	47.2	November	24.3
June	28.7	December	51.3

1. High value is due to snow melt.

As the chloride concentration in the rainwater is different from one rain to another, the average Cl_p must be taken, i.e.

$$Cl_p = \frac{Cl_1 \times P_1 + Cl_2 \times P_2 + Cl_3 \times P_3 \ldots Cl_n \times P_n}{P_1 + P_2 + P_3 \ldots + P_n}.$$

The chloride content can be obtained by using a bulk monthly sample.

Springs provide evidence of the existence of some natural resources, but only of shallow ground-water bodies and not underflows. There are two categories, those which circulate entirely within the zone of aeration, and those that are fed by a ground-water body.

Springs circulating within the zone of aeration are recharged by seepage water. Their discharge varies with rainfall. Each heavy rain causes a sudden, abundant flow of water through the large conduits, followed by a prolonged but much smaller flow through the narrow conduits. In such a case it is difficult to apply mathematical formulae to the flow from springs.

A study of the discharge curves can provide valuable information on springs down gradient from a saturated or waterlogged area (Fig. 14.3.5.3a).

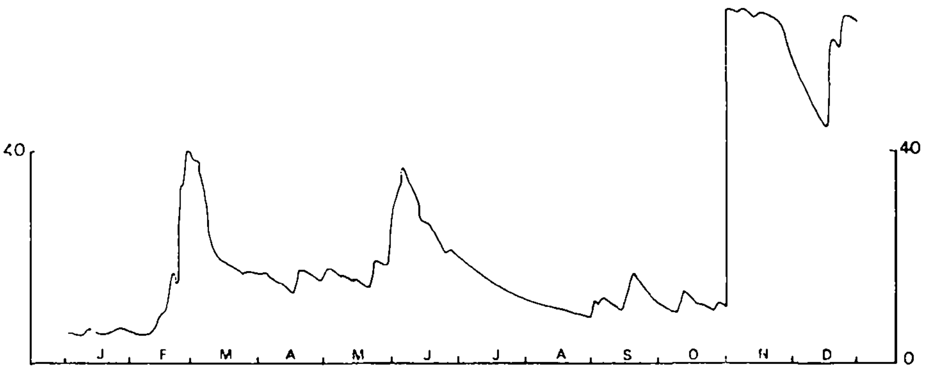

FIG. 14.3.5.3a. Discharge from the Fontaine de Vaucluse in 1968 (in m^3/s). (Flandrin and Paloc, 1969.)

During a rainy period, a ground-water body is recharged and, at the same time, discharges its water. In a prolonged dry period the aquifer is not being recharged but it empties itself as any reservoir would. The régime is said to be non-influent.

If the recharge rate exceeds the discharge capacity, the volume of ground water increases, being at its greatest at the time when the recharge becomes less than the discharge, and spring discharge is a maximum. There are three successive stages: (a) the recharge exceeds the discharge, and spring flow is each time increased by the amount of the excess; (b) there is still recharge, but at a rate less than the discharge. The spring flow diminishes to an amount equalling the sum of the drainage from storage and the recharge during the period; and (c) recharge stops, and the system becomes non-influent.

There is always a time-lag between the maximum recharge and the peak flow from the spring, which can last for several months. This time-lag is greater where the drainage capacity is low, as in the case of massifs which are poorly karstified.

On the other hand, when the ground-water body is able to discharge more water than it receives, the water flows straight through. There is a time-lag of days or hours between the peaks of rainfall and spring flow as happens when karstification is well developed. Conduits in which the discharge capacity greatly exceeds that of the intake are found in intensively karstified rocks, mountain karsts, old plains country and plateau karsts.

Other factors have to be considered. When surface karstification is poorly developed, time is needed for the water to infiltrate and there is no immediate increase in spring flow. Such may occur when a limestone massif which is resistant to karstification, perhaps because it is marly, rests on an easily karstified limestone, or as also is the case with a limestone massif overlaid by sandy sediments. Such conditions regulate the spring flow.

In a non-influent régime the variation of discharge from a spring occurs as follows. In the case of a ground-water reservoir with a horizontal impermeable base and very moderate thickness, for example, a perched water body, when Q_0 is the discharge at time t_0, Q_t the discharge at time t, and Q_p a permanent discharge of constant value, for laminar flow:

$$Q_t - Q_p = \frac{Q_0 - Q_p}{[1 + \alpha(t - t_0)]^2}. \qquad 14.3(15)$$

In the above formula

$$\alpha = \frac{a}{\Theta} \cdot \left(\frac{Q_0 K}{X^3}\right)^{1/2}, \quad \text{or} \quad \alpha = \frac{b}{\Theta^2} \cdot \frac{K V_0}{X^3}.$$

For turbulent flow:

$$Q_t - Q_p = \frac{Q_0 - Q_p}{[1 + \alpha(t - t_0)]^3} \qquad 14.3(16)$$

with $\alpha = a' \left(\frac{K}{\Theta^{3/2} X^2}\right)^{2/3} \cdot Q_0^{1/3}$ or $\alpha = b' \cdot \frac{K V_0^{\frac{1}{2}}}{\Theta^{3/2} X^2}$.

Where unconfined ground water is being discharged below the surface of another water body, or in the case of water movement within a confined aquifer, for laminar flow:

$$Q_t - Q_p = (Q_0 - Q_p)\, e^{-\alpha(t - t_0)} \qquad 14.3(17)$$

in which $\alpha = a'' \frac{K}{\Theta X}$ or $\alpha = a'' \frac{K h_0}{\Theta X^2}$,
and for turbulent flow:

$$Q_t = Q_0 - \alpha(t - t_0) \qquad 14.3(18)$$

in which $\alpha = b'' \frac{\varepsilon K^2}{\Theta X}$

where:

a and b = constants;
h_0 = the average thickness of the unconfined ground water;
ε = the thickness of a confined aquifer.

In equation 14.3(15), α is obtained by plotting $1/\sqrt{Q}$ against t on a graph; in equation 14.3(17) by plotting Q against t on semi-logarithmic paper, Q being on the logarithmic scale. α is known as the recession constant which increases in proportion to increases in the hydraulic conductivity K, the volume of water V_0 of the body in time t_0; and to decreases in the length of the ground-water body X, and the porosity θ.

Furthermore, if the area f of the spring outlet is constricted in comparison to the cross-sectional area F of the ground-water body, then a and b will decrease as the ratio f/F decreases. Therefore, a and b are direct functions of f/F.

The recession constant α depends both on the properties K and θ of the aquifer, and also on the volume of water V_0 or the average aquifer thickness V_0/X between intake and outlet. *The value of* α *can vary with the length of the path of travel.*

For springs discharging from highly karstified rocks, and generally having a high value for the variability index I, then α is > 0.01, sometimes as high as 0.9 (t being computed in days). I is the ratio of the greatest yield over the lowest one observed over many years. For springs discharging from poorly or non-karstified rocks having small values of I, then α is < 0.005.

TABLE 14.3.5.3a. Examples using formula 14.3(17)[1]

Spring	α_1	α_2	I
Highly karstified limestones			
Kef spring (Tunisia)	0.86–0.03	0.0065	16.5
Vaucluse spring (France)	0.10–0.007	0.01	27
Abime spring, Périgueux (France)	0.049	0.0026	11
	α		
Lez spring, Montpellier (France)	0.40		91
Kesra spring (Tunisia)	0.031–0.13		
Slightly karstified limestones			
Cerilly spring, Vanne (France)	0.0035		4.81
Miroir spring, Vanne (France)	0.00168		2.6
Dvuis spring, Vanne (France)	0.00127		1.7
Camblane spring, Gironde (France)	0.002–0.0016		2.2
Merignas spring, Gironde (France)	0.003–0.0015		

1. α_1 refers to large yields, α_2 to small yields.

These equations permit the calculation of the rate of discharge of a spring at the end of a lengthy drought, and also the total volume discharged during the drought, as follows. In the case of equation 14.3(17)

$$V_0 - V_t = \frac{Q_0 - Q_t}{\alpha}, \quad \text{or} \quad V_0 - V_t = \frac{\alpha}{2}(t^2 - t_0^2).$$

In the case of equation 14.3(18)

$$V_0 - V_t = \frac{Q_0{}^2 - Q_t{}^2}{2\alpha}$$

During a non-influent period, the level of the ground water falls, and its variation is given by the following equation:

$$H = H_0\, e^{-\alpha(t-t_0)}.$$

H_0 is the thickness of the ground water at time t_0, and H its thickness at time t. H_0 is difficult to calculate, but it is known that $\Delta = H_0 - H$: which leads to the equation:

$$\frac{\Delta}{\Delta_1} = \frac{1 - e^{-\alpha t}}{1 - e^{-\alpha t_1}}$$

from which $\log \Delta - \log \Delta_1 = \log(1 - e^{-\alpha t}) - \log(1 - e^{-\alpha_1 t}$
Δ, Δ_1, t, t_1, being known, α can be determined.
Two curves are plotted (Fig. 14.3.5.3b), one of $\log(1 - e^{-\alpha t})$ and the other of $\log(1 - e^{-\alpha t_1})$ against various arbitrary values of α. The value of α is sought at which the interval between the two curves is equal to the value of $\log \Delta - \log \Delta_1$. This α value is the one required to permit the calculation of H and H_0.

Water in storage. Reserves of stored water are essentially held in the zone of saturation, but in this regard it is important to know the amount stored per unit of aquifer volume, or the effective porosity, according to depth.

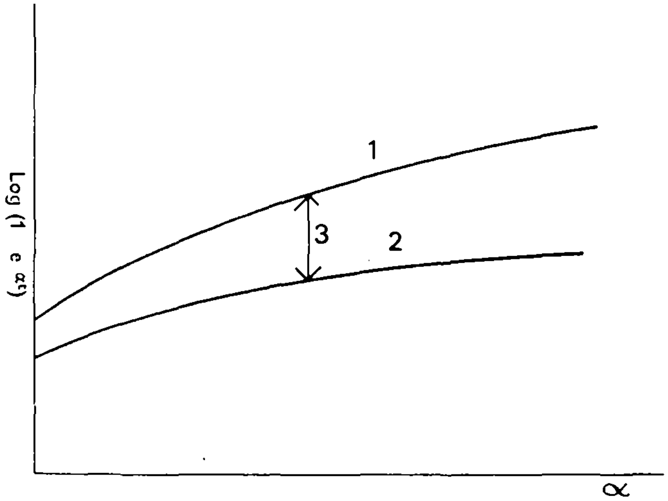

FIG. 14.3.5.3b. Determination of a. 1. log $(1 - e^{-\alpha t})$; 2. log $(1 - e^{-\alpha t_1})$; 3. log Δ − log Δ_1.

Near the top of an unconfined aquifer, the effective porosity can be more than 10, but it diminishes rapidly with depth. This high porosity is unequally distributed along the water-table, being higher along the main direction of ground-water flow. At greater depth, the porosity decreases rapidly, first to 1 per cent, then 0.1 per cent and even less where circulation becomes virtually non-existent.

This is true only for one type of limestone. Porosity is higher in heavily fractured and bedded limestones, and lower in massive limestones. In the upstream part of a confined aquifer, in its gradient part within a short distance of the outcrop, the lateral circulation has porosity characteristics similar to those in unconfined water. At a distance from the outcrops, the porosity is low, especially in an enclosed artesian system, because there is then no flow to enlarge the fissures. On the other hand, the porosity may be higher in an open artesian system, but this will be localized in the zones of active circulation. Two examples can be given.

1. Perched synclines. Because of their shape, perched synclines do not store the infiltrating water, which flows more or less rapidly according to the degree of karstification. As a result, during the rainy season the springs are well supplied and in rainless periods they tend to dry out. Water resources in this situation are represented by the volume of water available in storage at the beginning of the dry season, an amount which will be dissipated progressively by spring flow. This is the drainage curve which will make it possible to determine the method of using up resources and the volume available at the beginning of the dry season.

2. Buried structures. These include structures which are partly exposed at the surface, the remainder being buried to different depths. Under such conditions it is possible to store infiltrated water, resulting in the appearance of overfill springs, which drain off the excess. The amount of water which can be developed as a resource is equal to the volume of overflow together with what can be drawn from storage.

14.3.6 *Field investigations*

Here only the special aspects of investigation which apply to carbonate rock aquifers will be mentioned.

14.3.6.1 Geomorphological investigations

Geomorphological research, which can be carried out on foot or by vehicle, is greatly facilitated by observations from the air with the aid of aerial photographs.

Attention should be focused on: (a) topographical features peculiar to karst such as dolines, sinkholes, shafts, poljes; (b) condition of limestone surfaces, such as soil and vegetation cover, and the number and types of openings; (c) the physical condition of the limestones, that is whether compact, well-bedded, etc.; (d) presence of crush zones, fractures or faults; (e) the kinds of vegetation, which reflect the presence of ground water; (f) the topography.

14.3.6.2 Geological investigations

Stratigraphical and tectonic research and cartography will be carried out in the same way as in other terrains. In this case, however, it must be emphasized that cartography, in particular, must give priority attention to the lithology, even to the point of specifying the nature of the limestone in regard to the circulation of water. Obviously, a map will be of use only if the scale is at least 1/50,000.

14.3.6.3 Geophysical investigations

Geophysical investigations (Astier, 1971) can be helpful. Their nature will depend on what one is looking for, and on the existing hydrogeological conditions.

Investigation of the zone of aeration. These will be directed towards locating galleries, chambers and other cavities. Gravimetric or electrical prospecting may be useful, the latter for detecting the galleries near the surface which gravimetry cannot show, and gravimetry for indicating deeply buried cavities.

In the case of water-filled galleries, gravimetric and electrical prospecting can again be used in appropriate cases. Electromagnetic and seismic methods can also be used in the case of turbulent flow producing vibrations which reach the land surface, where this can be picked up by sound detectors. Where it is possible to reach the water at some point in the gallery, electric and electromagnetic methods with earthing can be applied. The flow path can thus be traced.

Investigations of the zone of saturation. It is important to find the surface of the zone of saturation. Electrical soundings will give the depth of the upper interface, and the depth of the lower interface between the zone of saturation and the limestone. Seismic refraction will do the same.

Infra-red radiometric prospecting. Karstic water infiltrating in high ground and moving to lower altitudes does not have time to warm up on its way downwards, and its temperature is commonly lower than the annual mean. Where water has long been in contact with the rock, especially if circulation is relatively slow, the emerging water has a temperature more or less the same as the mean annual air temperature. Thus in winter it will be warmer and in summer cooler than in the adjacent rocks. For this reason it can be detected if it is at shallow depths or if it moves towards the surface. Springs discharging into rivers or into the sea can also be located even if the temperature difference is small.

14.3.6.4 Use of tracers

Tracers can be used for: (a) determining whether or not there is direct communication between one point and another; (b) determining the direction of flow; and (c) measuring the velocity of ground-water movement and measuring yield.

In addition to the tracers already mentioned (7.4.2) for investigations of fractured and fissured rocks, it is also possible to use substances such as, in particular, oat-chaff, starch grains which are easily detectable with iodine, selaginella spores, yeasts, bacteria. These have the disadvantage of sometimes adhering to the walls of channels or small fissures.

Soluble chemical tracers, such as chlorides ($NaCl$, $CaCl_2$, $LiCl$, NH_4Cl), have to be diluted and used in large volumes to prevent them from sinking and escaping from circulation because of their high density. Their use is costly, and it may be better to use potassium bichromate, which even in a dilution of 1 to 2×10^{-9} can be detected by a diphenylcarbazide reagent. Phenol transformed into chlorophenol can be detected either by smelling or tasting.

Tracer dyes. Fluorescein, rhodamine and uranine in concentrations of 1×10^{-9} can be detected with the aid of a fluoroscope, and also with the aid of traps of active carbon which are made to surrender the absorbed tracer. The required quantity of fluorescein is given by the formula:

$$W = 0.5 \frac{q\,L}{v} + 0.02\,V$$

where:

W = weight in kg of fluorescein;
q = discharge of springs in m^3/s;
L = distance of travel;
v = velocity in m/day of the underground water;
V = volume of stored water in thousands of m^3/day.

Other dyes are eosin, erythrosin, Congo red, methylene blue, aniline blue, and Ponceau red (7.4.2). The quantities which have to be used are five to ten times those of fluorescein.

However, tracers are most readily transported along the quickest path of water movement, which the field observations may not truly measure unless the injection and observation wells are correctly placed. Reliability can be improved by using a large number of observation wells.

14.3.6.5 Use of boreholes and wells

These should not be used in the zone of aeration, but must penetrate saturated permeable rocks from which pumping will draw water from a wide surrounding area.

In the case of unconfined ground water, the best areas are those with a thick zone of saturation at shallow depth. The largest supplies of confined ground water are probably obtainable from the upstream part of the aquifer, where cavities are best developed. A borehole below a spring at the top end of a confined ground-water body will enable the spring to be captured and lateral leakage to be recovered, being then a new catchwork. Nearer the middle part of the aquifer, there may be less chance of finding good water-filled conduits. However, because limestone has a low plasticity and readily cracks when folded, drilling is sometimes worth while on anticlinal crests and along the axes of synclines. Finally, in an artesian basin there may be conduits in paleokarsts which should also be tested. Such paleokarsts may have been formed during a more or less protracted period of widespread uplifting of the land above sea level.

Exploratory boreholes not being used for production can be preserved as piezometric tubes for continued measurement of water levels. The methods used for test pumping are the same as those for rocks with interstitial porosity.

In relatively highly fissured limestones, pumping creates a cone of depression which is usually elliptical in shape because of the general anisotropy of limestone aquifers. The shape will be apparent from the piezometric wells.

It is to be noted that pumping can induce flow having a high Reynolds number in the vicinity of the well screen, either alone or together with a laminar movement. The relation then is $q = as + b\sqrt{s}$, in which s is the drawdown less the entry and friction head losses. Thus, in Tunisia, the equations for different boreholes (Gosselin and Schoeller, 1939) are as follows: Cedria borehole (Campanian), $q = 1.94 + 14.7\sqrt{s}$; Mareth borehole (Campanian), $q = 2 + 3.75\sqrt{s}$; West Assoued borehole (Maertrichtian), $q = 0.15 + 5.5\sqrt{s}$.

Where a steady-state flow can be assumed in a well-fissured limestone, the formula for aquifers with interstitial porosity can be applied. In steady-state flow in an anisotropic aquifer, the problem can be solved by changing the co-ordinates as follows:

$$X_1 = \frac{x}{K_x}, \quad Y_1 = \frac{y}{K_y}, \quad Z_1 = \frac{z}{K_z}.$$

For non-steady-state flow, see Papadopoulos (1967). For a steady-state turbulent flow the formulae are as follows:

For ground-water bodies or trenches:

$$Q^2 = K^2 \frac{h_1^3 - h_2^3}{3\,(r_2 - r_1)} = K^2 \frac{H^3 - h^3}{3\, r_e};$$

For perfect wells in an unconfined aquifer on an impervious horizon:

$$Q = 2\pi K \sqrt{\frac{r_e}{3}\,(H^3 - h^3)};$$

For a perfect borehole, in a confined aquifer:

$$q = 2\pi K m \sqrt{\frac{r_w\, r_e}{r_e - r_w}\, s},$$

and, as r_w can be neglected in comparison with r_e,

$$q = 2\pi\, K\, m\, \sqrt{r_w s}$$

where:

r_w = the radius of the borehole;
r_e = the radius of the drawdown cone of influence;
s = the drawdown in the well;
m = the thickness of the aquifer.

According to Hantush (1966) with unsteady laminar flow in an anisotropic aquifer, the drawdown s at time t with a constant rate of pumping Q, is given by the equation (Fig. 14.3.6.5a):

$$s = \frac{Q}{4\pi\, T_e}\, W\left(\frac{r^2}{4 \nu\, t}\right), \quad T_e = \sqrt{T_x\, T_y}$$

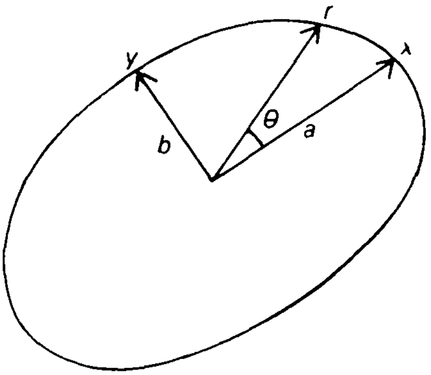

FIG. 14.3.6.5a. Unsteady flow in an anisotropic aquifer.

where:
T_e = average transmissivity of the aquifer;
T_x = highest transmissivity in direction of x;
T_y = lowest transmissivity in direction of y;
v' = Tr/S, the diffusivity coefficient of the anisotropic aquifer in direction of the observation well, S representing the storage coefficient.

Where r is equal to the distance from pumping well to observation well,

$$\frac{1}{T_r} = \frac{1}{T_x}\cos^2\Theta + \frac{1}{T_y}\sin^2\Theta;$$

$$T_r = \frac{r^2}{ab}\,T_e;$$

$$\frac{a}{b} = \sqrt{\frac{T_x}{T_y}} = \frac{T_c}{T_y} = \frac{T_x}{T_e}.$$

In the case of pumping for a long period:

$$\frac{t_0}{t'} > 100\,U_0, \quad U = \frac{r^2}{4at_0}.$$

Thus:

$$s = \frac{Q}{4\pi T_c}\log\mathrm{n}\,\frac{1+\Theta}{\tau}; \quad \tau = t_0/t_1.$$

Little is known about flows with high Reynolds numbers in anisotropic porous media, and the problem is being investigated. Analogue modelling is one of the modern tools for predicting water inflow to mines and galleries in karstic rocks (Babushkin *et al.*, 1972).

14.3.7 *Exploitation*

The present text concerns carbonate rocks, in which conditions differ from other rock types because water flow is mainly through fissures and channels in which head losses are small. Springs are common, and their discharge rates high, but subject to rapid fluctuation because the water circulates rapidly. These conditions differ from those in rocks with interstitial porosity where pulses of water intake rapidly flatten out and the springs react less sensitively. In carbonate rocks it is possible to exploit both springs and stored ground water.

14.3.7.1 Exploitation of springs

Spring discharges may vary from being very large after recharge periods to small during dry periods, and it is important to find a means of obtaining a constant and regular supply in such conditions.

Contact springs. It is not possible to increase the discharge or to regulate the flow from a simple flowing spring except by completely intercepting the water before it reaches the spring. Exploitation will be governed by the recession curve, having regard to the predicted minimum spring flow.

Overflow springs and overtopped springs. In these cases the ground-water body extends below the level of the spring. If a bore is drilled near the spring it is possible to obtain a constant supply by pumping from the bore at a sufficient rate to adequately augment the spring flow. As the flow decreases the pumping is proportionately increased.

During dry periods, pumpage will exceed the low-flow discharge, creating an empty space in the reservoir which will refill during the next period of high discharge. This will permit a constant rate of withdrawal approximating the average annual discharge of the spring, and an average yield markedly higher than the low-water discharge. With some overtopped springs, the aquifer can be reached through a gallery at a lower level, and a penstock in the gallery will enable the discharge to be regulated.

Depression springs. These are found in poljes, and their flows can be regulated by pumping.

14.3.7.2 Exploiting ground water in storage

Biological contamination of the ground water is most likely when there is no soil or sand cover to act as a filter. Contamination may also be caused by the incursion of saline water from the sea along a coastline, or from saline lakes or river flows.

Some ground-water bodies may be so far below the surface as to make their exploitation uneconomic. Where it is necessary to use them, one compensation is that periodic fluctuations in recharge do not cause adverse operating conditions.

It is usual to develop the major springs only, and much water may be lost through smaller springs around the edges of the structure. Development should always operate in such a manner as to reduce these losses to a minimum.

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15

Geochemistry of ground water[1]

15.1

Introduction

Section 7.4.4, entitled 'Observations of Ground-water Quality', already deals with types of analyses, pore solutions, methods of sampling and the presentation of chemical data. This supplementary chapter on ground-water quality will not repeat what has been given before, and reference should therefore be made to section 7.4.4.

In evaluating ground-water resources, a study of the water's chemical composition is almost as important as that of its movement, storage and recharge. Such a study is necessary to determine the suitability of the water for drinking purposes, for various domestic uses, agriculture, industry and others; it also provides valuable information about the recharge and its origin, the rate and direction of water movement, the discharge of ground water, its accumulation and storage, the nature of the aquifers through which it passes, and the possible presence of hidden mineral deposits. A study of the chemical composition of the water is vital to any assessment of the quality.

By 'quality of water' is meant its chemical, physical and biological characteristics, which may be excellent, good or poor.

The 'natural quality of water' is the quality of the water as it is naturally present in various rock terrains.

'Pollution' is the addition of chemical, physical or biological substances or of heat which cause deterioration in the natural quality, generally through the action of man or animals or any other kind of activity.

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[1] This chapter has been prepared by Dr H. J. Schoeller on behalf of the *ad hoc* subgroup of the IHD Working Group on Groundwater Studies and revised by a panel of editors composed of: A. A. Konoplyantsev, E. P. O'Driscoll, H. J. Schoeller and G. C. Taylor.

15.2

The general hydrochemistry of ground water

Ground-water quality cannot be fully understood without some background knowledge of general hydrochemistry. The chemical elements found in ground water come from rain, organic action in the soil and decomposition of the minerals in the rocks through which the ground water travels.

Some of these constituents are present in important quantities (1-1,000 mg/l); others are secondary, or minor, as shown below: (taken from Davis and De Wiest (1966)):

More important constituents (generally 10 to 1,000 mg/l): sodium, calcium, magnesium, bicarbonate, sulphate, chloride, silica.

Secondary constituents (generally 0.01 to 10 mg/l): iron, strontium, potassium, carbonate, nitrate, nitrite, fluoride, boron.

Minor constituents (0.0001 to 0.1 mg/l): *antimony*,[1] aluminium, arsenic, barium, bromide, *cadmium*, *chromium*, cobalt, copper, *germanium*, iodide, lead, lithium, manganese, molybdenum, nickel, phosphate, *rubidium*, selenium, *titanium*, uranium vanadium, zinc.

Trace constituents (generally less than 0.001 mg/l): beryllium, bismuth, *cerium*, cesium, gallium, gold, indium, lanthanum, *niobium*, platinum, radium, ruthenium, *scandium*, silver, *thallium*, *thorium*, tin, *tungsten*, ytterbium, *yttrium*, *zirconium*.

15.2.1 Rain-water

Rain-water itself contains chemical constituents, and although these are present in small quantities, at times they must be taken into account because of their subsequent concentration through evapotranspiration. In rain-water, Cl is present in concentrations from 0.1 to 100 mg/l, SO_4 from 0.1 to 20 mg/l, Ca from 0.1 to 20 mg/l, Mg from 0.1 to 10 mg/l, Na from 0.1 to 50 mg/l, etc. The highest concentrations of salts in rain-water are found between the coast and perhaps 50 km inland, as shown by the examples from Austria, Czechoslavakia and the Federal Republic of Germany given in Table 15.2.1.

TABLE 15.2.1. Examples of salts concentration in rain-water (mg/l)

Place	Distance from sea (km)	Salt			
		Ca	Mg	Na	Cl
Westerland	0.2	1.43	1.79	18.45	37.60
Schleswig	50	1.18	0.4	2.36	1.59
Braunschweig	450	2.74	0.52	0.74	2.12
Augustenberg	800	1.8	0.37	0.36	1.73
Hohenspeissenberg	900	1.56	0.27	0.19	0.80
Retz	1,200	1.91	0.22	0.16	1.28

---

[1] Constituents in italics occupy an uncertain position in the list. In practice, minor constituents are often associated with the trace elements.

Farther inland than 50 km, the continental influence can modify the relative concentration of the individual ions. Rain-water contains more salts in arid than in humid regions, because most dissolved salts are of continental origin, derived from salt-laden dust which has risen into the upper atmosphere. Near towns, there is an increase in SO_4^{--}, NO_3^-, etc., resulting from airborne pollutants.

15.2.2 *Chemical solution of minerals in rocks*

15.2.2.1 **Solution**

Solution affects soluble gases, and salts such as NaCl, KCl, $CaSO_4\ 2H_2O$, $CaCO_3$, $(CaMg)\ CO_3$. The solubilities of these constituents, expressed in g/kg of solution, are shown in Table 15.2.2.1a for three temperatures.

TABLE 15.2.2.1a. Solubilities of salts

t (°C)	NaCl	KCl	$CaSO_4$	$CaCO_3$	$(CaMg)\ CO_3$
10	263.6	238	1.926	—	—
20	264.4	255	2.020	6×10^{-3}	5×10^{-3}
30	265.5	272	2.095	—	—

Solution may involve two or more soluble substances at the same time, e.g. NaCl and $CaSO_4$. In addition, cations may enter the water through base exchange, so that new associations appear with the solubilities shown in Table 15.2.2.1b.

TABLE 15.2.2.1b. Solubilities of salts

t (°C)	$CaCl_2$	$MgCl_2$	Na_2SO_4	$MgSO_4$	Na_2CO_3	$NaHCO_3$
10	394	349	82.6	236	107	75.8
20	427	351	160.2	262	177	87.6
30	501	357	291.0	290	271	99.9

Concentrations may reach saturation point, and the solubility product then comes into play.
Given that:

$$\langle C^{Z_1} \rangle^{Z_1}\ \langle A^{Z_2} \rangle^{Z_2} = K,$$

where:
$\langle C \rangle$ and $\langle A \rangle$ = the activities of the cations and anions;
Z_1 and Z_2 = the valencies
the solubility product will be:

$$[C^{Z_1}]^{Z_1}\ [A^{Z_2}]^{Z_2} = \frac{K}{\gamma_c{}^{Z_1}\ \gamma_a{}^{Z_2}},$$

where:
$[C]$ and $[A]$ = the concentrations in moles;
γ_c and γ_a = the activity factors of C and A.

The activity factors vary with the ionic strength of the solution, that is, the quantity of various ions present. This leads to:

$$\mu = \tfrac{1}{2}\,[C_1Z_1^2 + C_2Z_2^2 \ldots + C_nZ^2{}_n],$$

where:
$C_1, C_2, C_3, \ldots C_n$ = the molal concentrations of the various ions;
$Z_1, Z_2, \ldots Z_n$ = the valencies of each ion.

As milligram equivalents, r, are generally used in hydrochemistry, this will give:

$$\mu = \left[\frac{rCl + rNa + rK + rHCO_3 + rNO_2}{2} + rSO_4 + rCO_3 + rCa + rMg + rFe^{++}\right] 10^{-3}.$$

Milligram equivalents (r) are obtained by dividing the weight of the constituent in milligrams by its atomic weight, itself divided by its valency.

For the commonest constituents, the atomic and molecular weights (p), the valencies (Z), and a factor (f) which converts the weight in milligrams per litre into milligram equivalents per litre are given in Table 15.2.2.1c.

TABLE 15.2.2.1c. Atomic and molecular weights, valencies and factors for the commonest constituents

	p	Z	f		p	Z	f
Na	22.9898	1	0.04350	HCO_3	61.02	1	0.01639
Ca	40.08	2	0.04990	SO_4	48.03	2	0.02083
Mg	24.312	2	0.08226	Cl	35.453	1	0.02821
Fe	55.847	2	0.03581	CO_3	60.00	2	0.03333
Fe		3	0.05372				
Sr	87.62	2	0.02283	NO_3	62.00	1	0.01613
K	39.102	1	0.02557	NO_2	46.01	1	0.2174
NH_4	18.04	1	0.05544	BO_2	42.81	1	0.02336
Al	26.9815	3	0.1112	Br	79.909	1	0.051251
Li	6.939	1	0.1441	F	18.9984	1	0.055264
Ba	137.34	2	0.01456	HPO_4	95.98	2	0.02034
Mn	54.938	2	0.0364	H_2PO_4	96.99	1	0.010310
Mn		4	0.07281	I	126.904	1	0.00788
Sr	87.62	2	0.02283	PO_4	94.98	3	0.03159

The relation of activity coefficients to ionic strength is given for important ions in the graph reproduced in Figure 15.2.2.1 (Hem, 1961).

In other cases the following formulae may be applied:

When $\mu < 0.01$,

$$-\log \gamma = AZ^2\sqrt{\mu},$$

where:
Z = the valency;
A = a temperature-dependent constant
(A = 0.489 at 0° C and 0.504 at 25° C);
when $\mu > 0.01$,

$$-\log \gamma = \frac{A\,Z^2\sqrt{\mu}}{1 + a\,10^8\,B\sqrt{\mu}},$$

where:

$$A = 0.4343\,\frac{\varepsilon^2}{2\,DRT}\,\sqrt{B};$$

$$B = \sqrt{\frac{8\,\pi\,N\,\varepsilon^2}{10^8\,DRT}}\,;$$

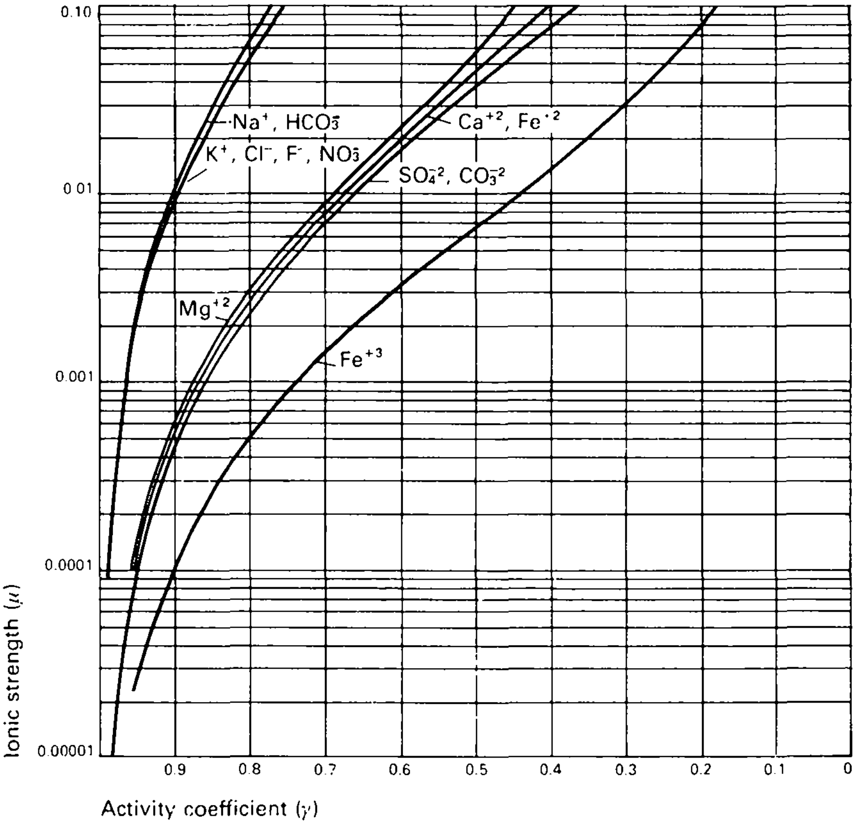

FIG. 15.2.2.1. Relation of activity coefficients for important ions in natural water to ionic strength of solutions (after Hem, 1961).

with:

N = 60161×10^3;

ε = 4.774×10^{-8};

D = the dielectric constant of the medium;

R = 8.316×10^7 cal °C^{-1} mole^{-1} (gas constant);

T = the absolute temperature;

$a\ 10^8$ = the effective diameter of the hydrated ions.

The values for A, B and $a\ 10^8$ will be found in Klotz (1964).

In addition to the simple ions, in some cases the pairing of some ions has to be considered, as in $CaSO^\circ_4$, $MgSO^\circ_4$, $MgHCO_3^+$, FeL^{++} (L being a ligand). The ionic pairs do not affect the solubility products of the simple ions. As a result, chemical analysis may give higher values than those indicated by the product of solubility.

15.2.2.2 Chemical attack[1]

Chemical attacks are the most important cause of the dissolution of minerals from rocks.

The carbon dioxide produced in soils by the decomposition of organic matter and by living organisms (micro-organisms, fungi, algae, roots and various animals) dissolves in water and produces carbonic acid. While the partial pressure of CO_2 in air is only 0.0003 atmospheres, it reaches 0.001 to 0.1 in soil.

Acids like HNO_3 and HNO_2 are produced by the decomposition of organic matter, and various other organic acids are produced by living matter.

Sulphuric acid is produced by the oxidation of sulphides such as pyrites. The most active agent is carbonic acid. The pH of water is an indication of its acidity or alkalinity, given that:

$$\text{pH} = \log \frac{1}{\langle H^+ \rangle}$$

At 24° C a neutral water has a pH of 7 as there are as many OH^- activities as H^+ activities.

$$\langle OH^- \rangle = \langle H^+ \rangle = 10^{-7}, \text{ when } \langle OH^- \rangle \langle H^+ \rangle = 10^{-14}.$$

When the pH value is less than 7, the water is acid, and when the pH value is greater than 7, the water is alkaline.

$\langle OH^- \rangle \langle H^+ \rangle$, however, varies with the temperature and so, therefore, does the neutral value of the pH, as shown in Table 15.2.2.2a.

TABLE 15.2.2.2a. Variations of neutral pH with temperature

°C	pH	°C	pH
0	7.47	24	7.00
5	7.37	25	6.98
10	7.27	30	6.88
15	7.18	40	6.69
20	7.08	50	5.45

---

[1] In this section, the angular brackets ⟨ ⟩ refer to the chemical activities and the square brackets [ ] to moles.

Thus water with a pH of 7 is acid at a temperature of 10° C and alkaline at a temperature of 40° C.

Limestone is dissolved when attacked by acids, the most important of which is carbonic acid. This attack is governed by equations expressing the balances between dissolved CO_2 and the HCO_3^-, CO_3^{--}, H^+, OH^-, Ca^{++} ions in the water, given in Section 14.3.4.1. These equations give the relationships between the values of the CO_2 of equilibrium and of the pH of equilibrium:

$$[CO_2]_{eq} = \frac{K_2'''}{K_1''' K_c'} [HC\bar{O}_3]^2 [Ca^{++}] \qquad 15.2(1)$$

$$pH_{eq} = \log \frac{K_c'}{K_2'''} - \log [HCO_3^-] - \log [Ca^{++}] \qquad 15.2(2)$$

$$pH_{act} = \log \frac{1}{K_1'''} + \log [HCO_3^-] - \log [CO_2] \qquad 15.2(3)$$

where:
pH_{act} = the measured pH value.

Equations 15.2(1) and 15.2(2) are valid when $pH < 8.3$ if CO_3^{--} is included as a part of HCO_3.

If the dissolved CO_2 content in the water is greater than the CO_2 of equilibrium, or if the pH of the water is lower than the pH of equilibrium, the water can still attack the limestone. In the reverse situation, the limestone precipitates.

The difference

$$SI_c = pH_{act} - pH_{eq} - \log_{10} \frac{\langle Ca^{++} \rangle \langle CO_3^{--} \rangle}{Kc}$$

is also called the Saturation Index of calcium carbonate. A negative value corresponds to an undersaturation, a positive one to a supersaturation.

The essential factor regulating the solution of limestone is the partial pressure of the CO_2. To this should be added the influences due to the ionic strength, μ, of the water and of the temperature.

Dolomitic limestone and dolomites are also attacked by carbonic acid (see Section 14.3.4.2.)

Silica and silicates: Silica is only slightly soluble in water. At pH values below 9 and at a temperature of 25° C, it occurs in true solution as $Si(OH)_4$ in a non-ionized state, but only when its concentration is less than 100-140 mg/l. At higher concentrations the $Si(OH)_4$ polymerizes and gives colloïdal gels. At pH values of less than 9, the pH has only a negligible influence on its solution.

When the pH value is greater than 9, the silica exists in the more or less dissociated state of $[SiO_4]^{4-}$ and its solubility becomes very high, rising from 200 to more than 2,000 mg/l of SiO_2 at $pH = 12$.

The temperature has a very considerable influence on the solubility of silicates. Table 15.2.2.2b shows the increase of solubility with temperature for a pH value of 7.5.

Silicates resist chemical attack, are slow to dissolve and contribute few constituents to water. All the ions comprising the silicates may dissolve in water. Only at a later stage may they recombine to form clays, kaolinite, montmorillonite, illite, etc., depending on the pH of the water and the nature of the ions.

Feldspar produces silica and alumina, Na, K, and Ca ions. Micas, amphibole and pyroxenes also produce silica, alumina, Ca, Mg and Fe ions, and olivine produces SiO_2,

TABLE 15.2.2.2b. Relation between solubility of SiO_2 and temperature

Temperature (° C)	SiO_2 (mg/l)	Temperature (° C)	SiO_2 (mg/l)
0	40	60	250
10	80	80	350
20	110	100	420
40	190		

Mg and Fe ions. Na, Ca and Mg ions mostly remain in solution; K ions are more or less adsorbed by the argillaceous products resulting from the recombination of silica and alumina. Fe precipitates, mostly through oxidation.

As silicates contain little Cl, $r\text{Na} + r\text{K}$ will always be greater than $r\text{Cl}$.

The relationship between the H_4SiO_4 content and the $\langle K^+\rangle / \langle H^+\rangle$ or $\langle Na^+\rangle / \langle H^+\rangle$ ratio depends on the nature of the minerals with which the silica is in equilibrium, and may thus indicate the minerals with which the water has been in contact (Figs. 15.22.2a, 15.22.2b).

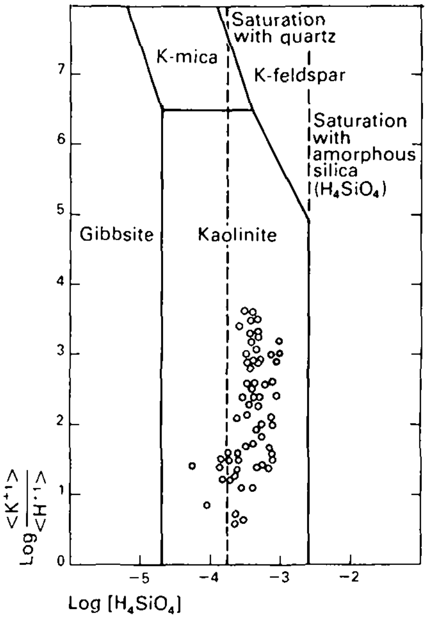

FIG. 15.2.2.2a. Stability relations of phases in the systems K_2O-Al_2O_3-SiO_2-H_2O at 25° C and 1 ATM total pressure as functions of $\langle K^+\rangle/\langle H^+\rangle$ and $\langle H_4SiO_4\rangle$. (After Feth, *et al.*)

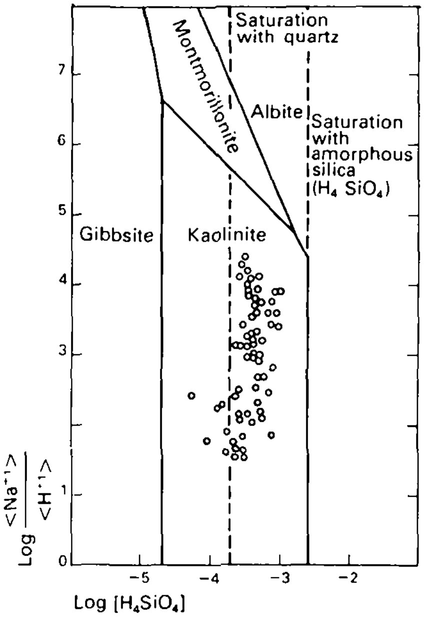

FIG. 15.2.2.2b. Stability relations of phases in the system Na_2O-Al_2O_3-SiO_2-H_2O at 25° C and 1 ATM total pressure as functions of $\langle Na^+\rangle/\langle H^+\rangle$ and $\langle H_4SiO_4\rangle$.

Generally speaking, the silica content of water depends on the minerals of the rocks through which the water travels, as shown in Table 15.2.2.2c.

TABLE 15.2.2.2c Relationship between SiO_2 content of water and the rocks percolated

Minerals	SiO_2 content of water (mg/l)	Minerals	SiO_2 content of water (mg/l)
Pure quartz	6-12	Shales	60
Amorphous silica	100-140	Feldspar, mica and other silicates, clays	12-100
Sandstone	60	Basalts	50

Iron, under ordinary conditions, is found in ground water in the form of Fe^{++}, $Fe(OH)_3$, or $Fe\,OH^{+}$ depending on the pH, the water's redox potential (E_h), and the HCO^{-}_{3} concentration. The iron content also depends on the pH and E_h values (Fig. 15.2.2.2c) and on its states of combination with organic matter.

15.2.3 *Changes in chemical composition*

15.2.3.1 Changes of equilibrium

Changes in chemical composition may come about through changes of equilibrium. The oxidation of organic matter such as peat, lignite, coal and hydrocarbons will produce CO_2, which increases the water's aggressiveness and hence its HCO_3 and Ca content. Conversely, the escape of CO_2 which occurs in caves or where there are springs, causes precipitation of $CaCO_3$.

A base exchange can bring about a considerable increase in Ca in a water that may be saturated with $CaSO_4$. The constant solubility product $[Ca^{++}]\,[SO_4^{=}]$ being reached, the addition of Ca will cause a decrease of SO_4. If the water is saturated with $CaCO_3$, or in other words if $[HCO_3]^2\,[Ca]$ is in equilibrium, this additional Ca will cause a decrease of HCO_3.

The dissolution of more salts such as NaCl increases the ionic strength of the water, and makes for a greater solubility of the other salts.

Changes of temperature also have an effect.

15.2.3.2 Reduction or oxidation

An increase of the oxygen content of the water increases the redox potential and causes oxidation of the iron, which will change from Fe^{++} to Fe^{+++} and be precipitated. The oxidized manganese will change from Mn^{++} to $Mn(OH)_2$, Mn_3O_4, Mn_2O_3, and will also be precipitated, the pH playing a part.

The oxidation of sulphides such as pyrites will add iron to the water, as well as sulphuric acid which reacts with the rocks.

Reduction of sulphates frequently occurs in the presence of organic matter, and leads to the production of S ions and consequently H_2S and HCO_3,

$$C + SO^{-}{}_{\bar{4}} + 2H^{+} \rightleftharpoons S^{=} + HCO_3{}^{--} + OH^{-}.$$

This reduction is sometimes so complete that SO_4 may disappear.

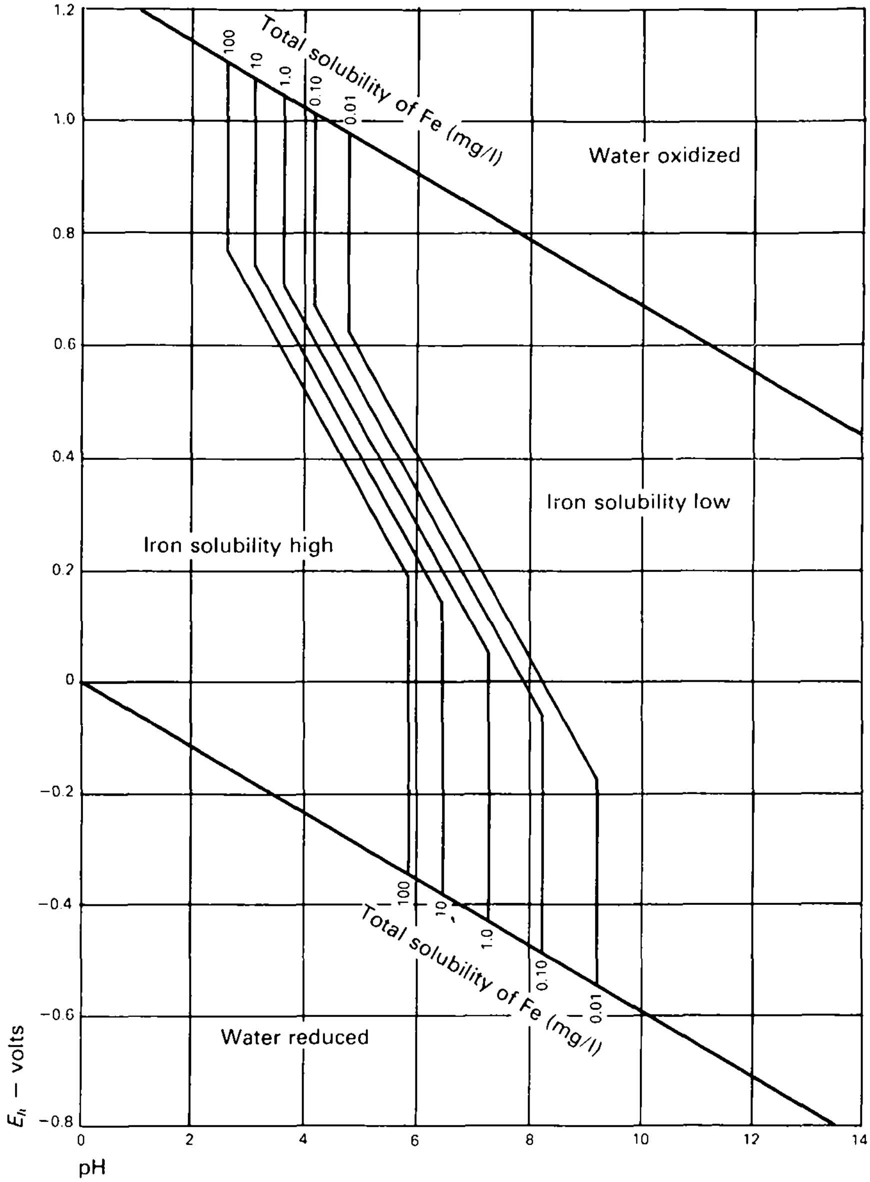

FIG. 15.2.2.2c. Solubility of iron in relation to pH and E_h (after Hem, 1961).

15.2.3.3 **Base exchange**

Argillaceous rocks and organic matter can absorb H^+, Na^+, K^+, Ca^{++}, and Mg^{++}ions. These are referred to as permutolites here because it is frequently impossible to identify them in an aquifer. Ion exchange may take place between these permutolites and the water.

$$2\ \text{perm. Na} + \text{Ca}^{++}\text{water} \rightleftharpoons \text{perm. Ca} + 2\text{Na water}$$
$$\text{perm. Mg} + \text{Ca}^{++}\text{water} \rightleftharpoons \text{perm. Mg} + \text{Ca}^{++}\text{water}$$

These exchanges cause a disequilibrium between the Cl ions and the Na and K ions. When water circulates in sediments it is usual to find $r\text{Cl} \approx r\text{Na} + r\text{K}$. A considerable difference between the concentrations of rCl and $r\text{Na} + r\text{K}$ indicates a base exchange represented by the base exchange index; (i):

$$i = \frac{r\text{Cl} - (r\text{Na} + r\text{K})}{r\text{Cl}} \quad \text{if } i \text{ is positive;}$$

$$i = \frac{r\text{Cl} - (r\text{Na} + r\text{K})}{r\text{HCO}_3 + r\text{CO}_3 + r\text{SO}_4 + r\text{NO}_3} \quad \text{if } i \text{ is negative.}$$

However, the breakdown of feldspars and other silicates gives $r\text{Cl} < (r\text{Na} + r\text{K})$. It is negative, and may be referred to as the index of inherent alkaline-chloride disequilibrium.

15.2.3.4 Concentration

Dissolved salts in ground water may become concentrated by solution processes as the water circulates, or by evaporation where the water table is at shallow depth, especially in arid regions.

The concentration of all constituents increases, except that of HCO^-_3, which depends on the CO_2 pressure, which remains constant. The $r\text{SO}_4/r\text{Cl}$, and $r\text{Mg}/r\text{Ca}$ ratios do not change at the beginning.

CaSO_4 saturation is first reached, and SO_4 can then increase only slightly because of the increase in the ionic strength of the solution. The $r\text{SO}_4/r\text{Cl}$ ratio decreases, but the $r\text{Mg}/r\text{Ca}$ ratio increases.

The sequence starts with:

$$r\text{HCO}_3 > (r\text{SO}_4 \text{ or } r\text{Cl}) \text{ and } r\text{Ca} > r\text{Mg} > r\text{Na} >$$

and is followed by:

$$r\text{Cl} > r\text{SO}_4 > r\text{HCO}_3 \text{ and } r\text{Na} > r\text{Mg} > r\text{Ca} >.$$

In the presence of very high concentrations this sequence may be modified through the addition of Ca by base exchange:

$$r\text{Cl} > r\text{SO}_4 > r\text{HCO}_3 \text{ and } r\text{Ca} > r\text{Mg} > r\text{Na}.$$

CaCl_2 and MgCl_2 are, in fact, much more soluble than NaCl.

15.2.3.5 Chemical filtration

When water under pressure is able to infiltrate through an argillaceous layer, the clay may act as a filter.

Cations will pass through if their diameter is sufficiently small, but anions are rejected. Uncharged water and H_2CO_3 molecules pass through. The result is that at the upstream side there is an increase in the concentration of Cl, Ca and Mg, caused by a loss of water. The water on the other side becomes more diluted, richer in HCO^-_3 and Na^+ and with a smaller proportion of Cl.

15.3 Water quality and natural conditions

15.3.1 *Influence of the rocks*

The nature of the rocks has an obvious influence on water quality, independently of climatic, ecological and topographical influence.

15.3.1.1 **Siliceous rocks**

Siliceous rocks such as sand and very pure sandstone contain few soluble substances. Water passing through them has a characteristically low dry residue (20–300 mg/l with an average of 220 mg/l, and a very low SO_4 and Cl content. The sequence is often $rNa > rCa > rMg$. The Ca content frequently varies between 4 and 60 mg/l, with an average value of 25 mg/l, and the HCO_3 content is about half of that which is normally found in water from other rocks, the usual values range from 12–160 mg/l with an average value of 60 mg/l, since the HCO_3 from carbonate is not added. Almost all the dissolved salts are introduced in rain-water and their concentration is due to evapotranspiration.

Where siliceous rocks contain even small proportions of limestone or clay, the water passing through them rapidly absorbs substances because there is a large area of contact. Depending on the degree of leaching and the intensity of evapotranspiration, the water may contain high concentrations of Cl, SO_4 and cations. The HCO_3 concentration remains normal.

15.3.1.2 **Carbonate rocks, dolomite and limestone**

In addition to the breakdown of the carbonate, other important factors influencing the salinity are the rate of circulation and the area of contact between the water and the rocks as well as the quantity of salts available for dissolution within the carbonate fabric.

Massive limestones contain mostly large fissures and channels and there are few narrow fissures. Water circulation is rapid, and the area of contact with the water is small in relation to the volume of water circulating. In addition, limestones of this kind are relatively pure and contain few soluble salts. To become dissolved in the water, these salts must pass through the compact limestone by diffusion, or be extracted by dissolution of the limestone. As infiltration is frequently rapid, evapotranspiration has only a small influence on the total concentration. The result is that the water has a very low salt content. The predominant constituents are HCO_3 and Ca, there being relatively little SO_4^{-}, Cl^{-} or Na^{+}. The Mg content depends on the amount of Mg in the carbonate.

When the limestones are porous or contain fine fissures, the water circulates slowly and the contact area between the water and the rocks is greater. Chlorides and sulphates can more readily be dissolved, especially from marly limestones. Under certain conditions, it may happen that $(SO_4^{-} + rCl) > rHCO^{-}_3$, the $rHCO_3$ remaining the same as before (Section 15.2.3.4), since it depends only on the normal CO_2 pressure.

Water from dolomitic limestone and dolomites has the same characteristics as water from limestone, but differs in its rMg/rCl ratio (Section 15.2.2.2).

The almost complete absence of filtration through fissures and channels makes water in carbonate rocks highly susceptible to pollution, particularly in organic and bacterial form.

15.3.1.3 Evaporites

Gypsum and anhydrite, or sulphated and salt-bearing rocks may have ground water passing through them. Because of the great solubility of these minerals, the ground water may contain large quantities of $CaSO_4$ and NaCl, even to the point of saturation.

Waters with all possible degrees of concentration may occur, depending on the rate of circulation, the distance of ground-water travel and the evaporite content of the aquifer fabric. HCO_3 remains normal except when the Ca concentration is large, in which case it may fall below the normal level. Generally speaking, HCO_3 is present in quantities smaller than $r\text{Cl} + r\text{SO}_4$ or than either taken separately. As the water becomes more concentrated, the $r\text{SO}_4/r\text{Cl}$ ratio decreases and the $r\text{Mg}/r\text{Ca}$ ratio increases.

Water which has leached a halite deposit will have a Cl/Br ratio greater than 300. Such water may reach the saturation point for $CaSO_4$ and occasionally for NaCl.

15.3.1.4 Carbonaceous rocks

Water which has been in contact with carbonaceous rocks, peat, lignite, coal and other hydrocarbons has a certain number of distinguishing characteristics: a negative redox potential often leading to an abundance of iron; frequent reduction of sulphates with a resulting decrease of the amount of SO_4; the possible presence of S^{--} and H_2S, and possibly an increase in HCO_3^-. When this water reaches the air, Fe^{++} is oxidized to Fe^{+++} and $Fe(OH)_3$, which precipitates.

The H_2S content does not adversely affect potability, unless present in large quantities, because it rapidly escapes into the air. However, its presence together with a negative redox potential and the presence of aggressive CO_2, make this water very corrosive to metal piping.

15.3.1.5 Argillaceous rocks

Water in clays and shales may be more or less salty depending on the degree of leaching. Clay also absorbs some ions. Water in contact with argillaceous rocks, or in aquifers containing clay, may become charged with salts. It may also undergo base exchanges: Na for Ca for Mg, or Ca for Mg for Na.

When the water circulates very slowly, as for example at a rate measured in centimetres per year, it may gain in Ca, Mg, Na, Cl and SO_4, and the base exchange process may become complete. In some cases, $CaCl_2$ and $MgCl_2$ appear as salts, and $r\text{Ca} + r\text{Mg}$ is greater than $r\text{SO}_4 + r\text{Cl}$, but in other cases Na_2CO_3 appears and $r\text{Na}$ is greater than $r\text{SO}_4 + r\text{Cl}$. When, on the other hand, rapidly circulating water has leached the rocks over a long geological period, these concentration and exchange phenomena are very slight.

15.3.1.6 Crystalline and magmatic rocks

In crystalline and magmatic rocks, such as granites and basalts, water circulates through fissures. In granite these are usually narrow, and are rarely deeper than 30 to 60 m. In basalt flows the fissures are wider and more numerous.

The silicates in these rocks are resistent to solution, so that under ordinary conditions, water from granite and basalt has a relatively small dry residue. The HCO_3 content is low and about half the normal for all rocks, because there is no additional HCO_3 coming from the rock. Such water also normally contains little Cl and SO_4, which mostly come in the rain-water. Any higher concentrations are the result of evapotranspiration. If the rocks are rich in sulphides, such as pyrites, their oxidation gives rise to SO_4 ions.

The Ca/Na ratio is often lower in granites than in other rocks, and an occasional tendency for Na to predominate over the other cations depends essentially on the nature of the feldspars. One of the characteristics of this water is that rCl is less than rNa + rK, which is an indication of an inherent negative alkaline chloride disequilibrium, not to be confused with base exchange.

As no HCO_3 ions are dissolved from these rocks, the HCO_3/free CO_2 ratio is low, which gives the water (Equation 15.2(3)) a low pH value of 4–6. When the water reaches the air, the CO_2 escapes, the pH rises, and the water may become alkaline through the predominance of Na. Iron from ferromagnesian minerals or pyrites then precipitates. As most frequently, the rNa + rCl and even rNa alone are greater than rSO$_4$ + rCl, the water is of the sodium sulphate or even of the sodium bicarbonate type.

Water from basalt is frequently distinguishable by a higher rMg/rCa ratio, the Mg being derived from olivine.

A high content of dissolved substances in the water is a result of long stagnation in fissures, or of concentration by evapotranspiration.

15.3.1.7 Metamorphic rocks

Water circulating in gneiss has the same characteristics as water circulating in granite.

Water from slates and schists generally has a higher dissolved content than water in crystalline and magmatic rocks. It frequently contains abundant SO_4 and Fe, resulting from the oxidation of pyrites.

Table 15.3.1.7 gives the typical chemical composition of waters of various origins.

15.3.2 *Influence of hydrogeological conditions*

Water quality is influenced not only by general physical and chemical phenomena and the petrographic nature of the aquifers but also by hydrogeological conditions.

15.3.2.1 Soil as the principal factor affecting composition

The principal rock dissolving agents are the carbon dioxide produced in the soil and the various organic acids originating from biological activity. In the zone of aeration and within the ground water itself, dissolution phenomena occur much less readily.

Dissolution occurs mainly in the soil, the constituents of the soil itself being merely the residues of the weathering phenomena which have affected the original rock, from which other substances have been dissolved. The constituents dissolved in this way are added to those from the rain-water. A large portion of infiltrated rain-water is removed through evapotranspiration from the soil, leading to a concentration of constituents in the water infiltrating to the water table.

As a result there is little noticeable difference between the chemical composition of the water which has percolated from the surface of the soil and that of the ground water at greater depth.

15.3.2.2 Effect of the distance travelled by the water

Further changes occur while the ground water is in transit through the rocks, although these changes are far less intense. As the water commonly travels a long way and takes a

TABLE 15.3.1.7. Typical chemical composition of waters (mg/l)[1]

	Place[2]	Total dissolved salts	Ca	Mg	Na	K	Cl	SO_4	HCO_3	Fe
Granite	1	122	14	4.7	22		5.9	52	46	
	2	83	7	2	18		17	4	24	0.1
	3	470	62	28	15		66	14	224	
Basalt	4	169	27	10	13		6.2	11	128	0.08
	5	165	15	8.9	11	3.0	5.0	2.6	109	0.01
	6	367	48	31	15	3.8	4.0	110	192	1.97
Calcarenites and others	7	437	107	12	42		61	36	355	
	8	540	132	25	34		72	101	355	
	9	884	157	38	99		171	245	342	
Compact crystalline	10	520	95	61	6		10	61	402	
	11	330	86	13	16		3.5	45	240	
Limestone	12	210	44	4.3	8		11	27	100	
Dolomites	13	890	103	82	83		141	303	336	
	14	440	68	46	29		90	58	147	
	15	662	87	56	68		142	207	192	
Shales	16	3,255	512	22	47		73	1,485	81	
	17	2,236	235	138	394		936	282	504	
	18	1,667	145	106	304		583	353	342	
Sandstone and pure sands	19	143	35	11	14		14	6.4	50	0.5
	20	189	41	3.4	5		23	26	63	0.2
	21	140	10	8	25		36	45	18	7.0
Sands and ordinary sandstones	22	482	59	37	67		74	84	336	
	23	439	65	38	44		63	79	326	
	24	504	69	27	69		77	162	200	0.4
Evaporites	25	5,609	536	601	315		650	3,368	204	
	26	2,881	613	126	72		98	1,887	150	
	27	245,300	1,071	1,974	88,897		137,788	11,981	324	
Lignites	28	1,208	16	6.6	585		36	11	1,506	0.1
	28	472	7	8	13		20	16	200	7

1. The chemical composition of the water from each of these rocks may vary considerably from the typical composition shown, depending on hydraulic or weather conditions and the presence or absence of soil, etc.
2. 1, Butte, Montana, United States. 2, France. 3, San Diego, California, United States. 4, Mud Lake Region, Idaho, United States. 5, 6, The Dalles, Oregon, United States. 7, 8, 9, Tunisia. 10, 11, 12, France. 13, 14, 15, Tunisia. 16, France. 17, 18, Tunisia. 19, 20, France. 21, 22, 23, Tunisia. 24, France. 25, 26, 27, Tunisia. 28, France.

long time to move through the system, there may be a considerable secondary concentration of constituents, particularly when the aquifer has not previously been strongly leached.

Especially in confined ground water, circulating slowly through the rocks where there is little leaching, there may be a general increase in the total amount of dissolved salts, with consequent changes in the chemical composition: HCO_3 remains constant, SO_4 and Cl increase while the rSO_4/rCl ratio remains the same. Once the total concentration of salts reaches a certain value, the rSO_4/rCl ratio will diminish, with an accompanying increase in the rMg/rCa ratio.

Finally, the sequence becomes $rCl > rSO_4 > rHCO_3$ and $rNa > rMg > rCa$, with the possibility, at the highest concentration, of the $(rCa + rMg) / (rSO_4 + rHCO_3)$ ratio being greater than unity. This change can be followed by establishing isocones (lines of equal concentration of total salts), isochlors, and lines of equal values of rSO_4/rCl, rMg/rCa, etc.

Only the isochlors can be used to evaluate the circulation of the water, since this is the only constituent which does not react appreciably with the rocks. The isochlors are distributed more or less in accordance with the following equation (M. Schoeller, 1964) (see Fig. 15.3.2.2a).

$$x = \frac{v}{a\,(C_m + b)} \cdot \log_e \left[\frac{C_m - C_o}{C_o + b} \cdot \frac{C + b}{C_m - b} \right]$$

where:

C and C_o are the concentrations at distance x and x_o;
C_m is the maximum possible concentration;
v is the velocity of the water;
a is a factor for the solution of salts from the rocks;
b is a constant which may be zero.

The curve $C = f(x)$ commences with an increasing dC/dx corresponding to the upstream portion of the aquifer where there is considerable leaching, and a second portion where dC/dx decreases towards the downstream part where leaching is at a minimum. The inflection point is at $(C_m - b)/2$.

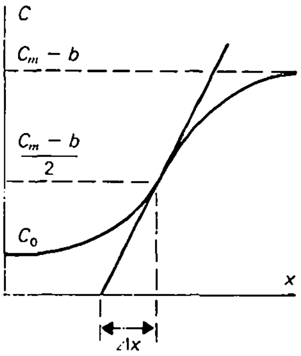

FIG. 15.3.2.2a. Concentration (C) by dissolution as a function of distance x. The point of inflexion is at $(C_m - b)/2$ (after M. Schoeller, 1964).

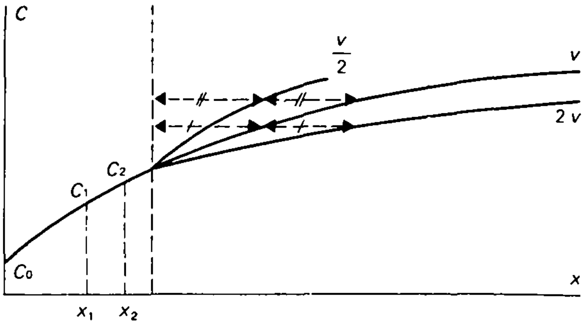

FIG. 15.3.2.2b. Salt concentration C in water by dissolution (influence of velocity) (after M. Schoeller, 1964).

Where the aquifer has not been leached, the curve becomes simpler (see Fig. 15.3.2.2b) so that

$$x = \frac{v}{a} \log_e \frac{C_m - C_o}{C_m - C}$$

Similar phenomena can be found in unconfined ground water, with an increase in concentration between the up-gradient and down-gradient areas.

When unconfined ground water is recharged along the whole of its path, and the rocks are completely leached, no more salts are taken into solution. The equation is then:

$$C = \frac{C_o Q_o + p C_p x}{Q_o + (p - e)x}$$

where:

C and C_o = the chloride concentrations in the ground water at distances x and x_o;
Q_o = the initial ground-water discharge;
p = the rainfall;
e = the quantity of water lost by evapotranspiration;
C_p = the chlorine concentration of the rainfall.

As the ground water moves downstream, its concentration may increase (curve 2, in Fig. 15.3.2.2c), remain constant or decrease (curve 1 of the same figure) depending upon whether C_i, the concentration of the water infiltrations down to the water table, is greater than, equal to, or less than C_o. The value of C_i is given by:

$$C_i = \left(\frac{p}{p - e}\right) C_p$$

When the unconfined ground-water surface approaches the soil surface, evapotranspiration may remove a large quantity of water, particularly in hot arid regions. As a result, a high salt concentration occurs which increases downstream. If, as is the case in hot arid regions, e is greater than p, the gradient of the curve of $C = f(x)$ will increase in the downstream direction (see Fig. 15.3.2.2d).

15.3.2.3 **Effect of the depth of the aquifer**

The deeper the ground water lies, the more slowly it circulates and the more difficult it is for it to be replaced by other water (particularly water infiltrating from the surface). Mineral composition, permeability and potentiometric gradients being equal, deeper rocks are less subject to leaching, and retain their water for a longer time. As a result, the dissolved salt concentration in the water tends to increase with depth. When there are no large evapotranspiration losses to affect the process, it is natural for the found-water composition to change from a bicarbonate type at shallow depth to a chloride type as the depth increases. The modification of the chemical composition with depth follows the pattern:

$$HCO^-_3 \rightarrow HCO^-_3, SO^=_4 \rightarrow SO^=_4, Cl^- \rightarrow Cl, SO^=_4 \rightarrow Cl^-$$

This pattern merely describes the variation in chemical composition which results from concentration. It leads to a certain vertical zoning, but the above pattern, however, like the continual increase in concentration, may not always hold. A permeable aquifer having an abundant inflow and greater rate of circulation may be interbedded in the series of strata. In this case, the water in this aquifer may be of a lower salinity than

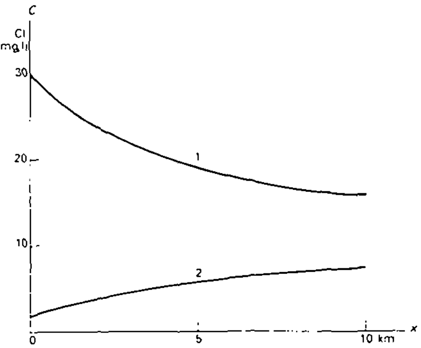

FIG. 15.3.2.2c. Influence of evapotranspiration on salt concentration in an unconfined aquifer in non-arid regions. $Q = 0.05$ l/s; $p - e = 250$ mm/year $= 8 \times 10^{-6}$ l/s.
1. $C_o = 30$ mg/l; $C_i = 5$ mg/l;
2. $C_o = 2$ mg/l; $C_i = 10$ mg/l.

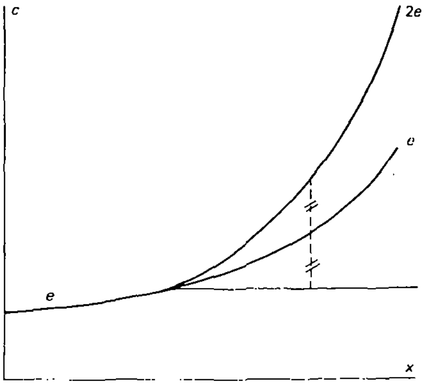

FIG. 15.3.2.2d. Influence of evapotranspiration on salt concentration in an unconfined aquifer in arid regions (after M. Schoeller, 1961).

that in the strata above and may, for example, show an HCO_3, SO_4 sequence beneath an SO_4, Cl sequence. It may also happen, particularly in arid regions, that the body of ground water nearest the surface, although receiving relatively fresh water, has a high salt content because of intense evapotranspiration.

15.3.2.4 Importance of paleohydrogeology

The chemical composition of ground water depends on paleohydrogeology. The composition is frequently inherited from past ages. Hydrogeological history has followed a series of cycles, each of which begins with an expulsion phase (elision phase, Kartzev, 1969) during sedimentation. During this phase the sediments are compressed and the water which they contain is expelled. This phase is followed by another during which the beds are infiltrated by rain-water, which may progressively displace the water of sedimentation. Each of these phases may last for a complete geological period amounting to millions of years. The dewatering of a stratum may continue after the first cycle, and further sedimentation may add a new load of superincumbent rock, causing the expulsion of more water.

Leaching during the infiltration phase may be considerable or slight, depending on the duration of the phase, the permeability of the rocks, and the rate of movement of the water. The water of sedimentation expelled from argillaceous rocks may penetrate into adjacent layers, and this process controls the geochemistry of ground water. For example, an ancient limestone may have remained above the water table for a long geological period and have become intensely karstified. Younger clay deposits subsequently covered the limestone. In the first phase, water squeezed out of clay deposits may have entered the limestone, but because of the karstification, the water was able to circulate freely. In this way original saline water was diluted and displaced, and only fresh water now circulates.

15.3.2.5 Connate and fossil water

There is no universally accepted precise meaning for either of these terms, because they may include waters of slightly different origins. In this text the term connate will be used for waters enclosed in a non-dynamic system, and which have had no contact with the atmosphere for a long period. Such water does not circulate under the influence of gravity, but is motionless. It originates as water entrapped in a sediment during deposition, although the present salinity may differ from that of the original water of deposition because of the subsequent effects of such processes as osmosis and diffusion. Connate water is usually, but not always too saline for use even by the most salt-tolerant plants or animals.

The case of water which has entered an aquifer forming part of a dynamic system is different. The rate of movement of the water through the system is so slow, perhaps only a few metres a year, that several tens of thousands of years may be required for a quantum of water to move completely through the system. Such water is commonly referred to as fossil water. For example, fresh water stored in the Nubian sandstone aquifers of the Sahara during the Pleistocene epoch is often called fossil water, even though it is part of a dynamic system.

15.3.2.6 Proximity to sea or salt lakes

When a fresh-water aquifer is connected to a surface body of salt water, such as the sea or a salt lake, the denser salt water will tend to penetrate some distance into the aquifer until a state of balance with outflowing fresh water is established. If the aquifer has an impermeable floor at a relatively shallow depth, the salt water will move over this floor as a wedge-shaped intrusion. The distance it penetrates into the aquifer will be inversely proportional to the rate of fresh ground-water discharge. Any decrease in the discharge or the pressure head within the aquifer, which may be caused by pumping, will cause the salt water to intrude farther into the aquifer, and may ultimately cause saline contamination of the pumped water. The fresh-water salt-water interface will advance or recede in accordance with variations in head caused by changes in the rate of pumping, or changes in atmospheric pressure, or under the influence of tidal fluctuations in the surface-water body. For a short distance inland from the shoreline, shallow unconfined ground water may also be contaminated by salt from airborne seaspray, especially during storms.

In shoreline karsts the hydrodynamic relations between sea and fresh water may be complex. An intruding wedge of salt water may penetrate far enough inland to contaminate the ground water discharging from coastal springs, the degree of contamination being determinable from the increase in salt concentration and the variation in the proportions of the dominant ions. For example, compared with the fresher ground water, the more saline mixture will have higher SO_4/Cl and Mg/Ca ratios, and a smaller proportion of HCO_3. The relative amounts of the various ions cannot be directly used to calculate the proportion of admixed sea water because of the base exchange reactions which take place. The Cl ion can be used for this purpose however.

15.3.2.7 Proximity to streams

Streams can affect ground-water quality in two ways. When a stream rises and floods, the higher water level will cause river water to penetrate into the ground water through permeable sections of the river bed and banks. The flood water may also inundate the adjacent river flats and raise the ground water level by downward infiltration.

If the ground-water level beneath or adjacent to a river is permanently lower than river level, the river loses some of its water, and modifications occur in the ground-water quality. This occurs commonly in fissured and karstic terrains.

15.3.2.8 Fluctuations in ground-water level

Water-table fluctuations may affect the quality of the water. If the ground-water body is at shallow depth, a larger number of plant roots may draw water from it when the water table rises. Transpiration and possibly evaporation losses will be greater, and the salt content will increase. Conversely, a lowering of the water table may induce lateral inflow from nearby areas, with a consequent change in salinity.

15.3.2.9 Chemical aureoles around mineral deposits

Mineral deposits may be above or below the water table. In general the host rock will in some measure be permeable, and circulating water may take some of the mineral into solution. This may not be important unless the mineral is toxic or in some other way undesirable, or if the ground water is not being used.

Under natural conditions the mineralized zone may come to be surrounded by an aureole of contaminated water, the chemical concentration of which decreases with distance from the mineral body as the water itself tends slowly to move radially outward and away. Trouble may arise, however, if the water has a total salinity low enough to make it attractive for use as, for example, drinking water. Pumping from a well or borehole will induce a more rapid movement of water towards the borehole. As a result water from the zone of contamination near a mineral body may gradually be drawn towards and finally reach the well. The pumped water will then be increasingly contaminated. Two recent Australian examples illustrate this condition. In the first, water from a proposed town water-supply borehole drilled in jointed granulite some distance from a copper lode gradually increased in copper content until it became unfit for use. In the other case, a diamond drill crew prospecting for lead had drilled a water-supply borehole in alluvium a short distance away from a lead lode and soon were hospitalized with lead poisoning.

15.3.3 Influence of climate

Salt concentration in ground water commonly takes place through evaporation and transpiration processes, and the chemical composition of the water may change according to the degree of concentration. The effect of evaporation will be greatest where the air is warmer, drier and more mobile. On the other hand, transpiration losses depend on the density of the vegetative cover, which in turn is dependent on the rainfall. In fact, the chemical composition of ground water depends to a large extent upon the soil and the vegetation which it supports.

Rainfall plays an important role too. Its intensity is important since downward percolation of salt from infrequent but heavy downpours is much greater than from prolonged but light showers. Rain-water may leach out soluble salts such as NaCl or $CaSO_4$, and carry them away to the streams, or it may leach surface accumulations of salts into the ground water. $CaCO_3$ behaves differently because, being only slightly soluble, it is dependent on the free CO_2 content of the water, a factor which is almost completely independent of the intensity of the rainfall. In the tundra regions of northern Europe the evapotranspiration is slight, and its action does no more than reduce the effect of the leaching process. In addition, little CO_2 is released by natural processes because the temperature is low, and the water is of the $rSiO_2 > rHCO_3^-$ type with low salinity, the total of dissolved salts being only 100 - 150 mg/l.

In a wet subtropical climate the temperature is higher, the production of CO_2 in the soil is more abundant, and evapotranspiration greater. Leaching occurs during the rainy season, and the climate produces podzols and leached brown soils. The total dissolved salts usually vary from 100 to 1,000 mg/l according to the nature of the rocks. Generally the sequence is:

$$r\mathrm{HCO_3} \text{ and } r\mathrm{Ca} > r\mathrm{Mg} > r\mathrm{Na}$$

Still nearer the equator, in a hotter, more arid climate, the rainfall is less than the evapotranspiration. The chlorides and sulphates, being less subject to leaching, will become concentrated in the soil by evaporation and subsequently flushed downward into the ground water after heavy rain. The passage of this additional water through the soil brings about a precipitation of $CaCO_3$, because the $CaCO_3$ content in the water is limited by the pressure of the CO_2 in the soil.

As a result of the greater concentration of the total amount of salts, the sequence becomes:

$$(r\mathrm{SO_4} \text{ or } r\mathrm{Cl}) > r\mathrm{HCO_3} \text{ and } r\mathrm{Ca} > r\mathrm{Mg} > r\mathrm{Na}$$

In arid regions the salt concentration is still higher because evaporation is greater, the rainfall lower, and downward percolation small. There is greater precipitation of $CaCO_3$ in the soil, and there may even be precipitation of gypsum, the sequence being:

$$r\mathrm{SO_4} > r\mathrm{Cl} > r\mathrm{HCO_3} \text{ and } r\mathrm{Ca} > r\mathrm{Mg} \text{ or } r\mathrm{Na}$$

or

$$r\mathrm{Cl} > r\mathrm{SO_4} > r\mathrm{HCO_3} \text{ and } r\mathrm{Na} > r\mathrm{Ca} > r\mathrm{Mg}.$$

In some regions the heavy tropical rainfall may be greater than the evapotranspiration losses, in which case the ground-water salinity will be low, the main anion often being HCO_3. In such a climate iron, aluminium and silica play an important role.

15.3.4 *Hydrochemical zones*

Independently of any geological factors, an initial subdivision into hydrochemical zones is broadly possible on the basis of climate, which itself varies with the geography, and exercises a dominant control over ground water hydrochemistry. The particular effect of this climatic control is that variations in salinity are frequently associated with variations in the ratios of elements one to another.

The climatic zonation can be subdivided on a geological basis because the chemical composition also depends on the rock type. For example, there are zones of igneous and metamorphic rocks, sediments, evaporites, limestones and pure siliceous sands. However, the effect of the geological environment is largely subordinate to that of climate.

On a more limited scale but perhaps even more importantly, ground-water salinities often increase with depth, because the rate of circulation decreases. This is well recognized in some deep sedimentary basins in Australia, (Miles, 1952, O'Driscoll, 1960) and also in the western part of the U.S.S.R. (Il'in, 1930; Posokhov, 1969).

15.4 Chemical and biological changes in water quality caused by artificial conditions

Biological pollution of ground water is not so marked as that of surface water, because the soil is a fairly effective filter between the source of pollution and the water table. For this reason, it is sometimes overlooked and serious cases of pollution may pass unnoticed, even though they do in fact occur.

15.4.1 *Pollution*

Before reaching their destination, pollutants may travel some distance over the surface, pass vertically downward through the soil and the zone of aeration as far as the water table, and then move laterally in the ground water.

15.4.1.1 **Filtration**

Filtration here means the removal of suspended impurities or pollutants from water during its passage through a relatively fine porous medium such as soil.

15.4.1.2 **Soluble substances**

Soluble substances may be mineral or organic, and are easily transportable by water merely because they are soluble. Some originate from domestic waste water and fertilizers, others from industrial or mining waste. Some contaminants are extremely toxic.

Although many metals remain soluble in a reducing medium, oxidation may cause them to precipitate from solution and be retained in the rocks as mineral deposits. They may also be retained by adsorption in argillaceous rocks or organic matter. Lastly, they may react with minerals in the rocks. Acidity, for example, may be neutralized by carbonate rocks.

15.4.1.3 **Non-soluble liquids**

Substances such as hydrocarbons may be in part retained in a host rock and in part flushed out by water, which has a higher surface tension. An unsaturated sand may retain a volume of hydrocarbon equal to 8 to 10 per cent of the total porosity.

If the sources of pollution overlies a ground-water body, the contaminants descend through the zone of aeration until they reach the upper part of the zone of saturation over which they spread. If the source is within the ground-water body, the less dense hydrocarbons may rise and also spread through the upper part of the zone of saturation and thence move down the hydraulic gradient.

15.4.1.4 **Solids**

Solids are transported by ground water only if the openings in the aquifers are sufficiently large, as in fissured rocks and karst areas.

15.4.1.5 Micro-organisms, viruses, pathogens

In addition to the rate and direction of movement of the ground water, transportation of these substances depends on the size of the pores, on reactions, on the amount of food available and on their life-span.

Micro-organisms are carried by ground water but tend to attach themselves by adsorption to the surfaces of solid particles. It is generally accepted that the maximum rate of travel of bacterial pollution is about two-thirds that of the water moving through rocks with interstitial porosity. However, the distance travelled by bacteria from the source of pollution rarely exceeds 33 m in such rocks, even though pathogenic bacteria may survive as long as 4 years.

In cavernous or fissured rocks the travel distance may be measurable in kilometres. Pollution of this type becomes important, however, only if the water is to be used for potable water supply or food processing.

Viruses which retain all their characteristics for more than 50 days may spread 250 m. or more in soils where there is organic matter to serve as a source of nutriment. The recommended safe distances between domestic wells and sources of pollution in non-karstic terrains are indicated in Table 15.4.1.5 (Romero, 1970).

TABLE 15.4.1.5. Recommended safe distance between domestic wells and sources of pollution

Source of pollution	Distance (m)	Source of pollution	Distance (m)
Septic tanks	15	Sewage farms	30
Pipes with watertight joints	3	Infiltration ditches	30
Other pipes	15	Dry wells	15
Percolation zones	30	Cesspits	45

15.4.1.6 Transport

Transport at the surface will not be discussed. Infiltration takes place vertically through the uppermost layer of the soils, a medium which has an enormous purifying power, not only because of its content of organic and dispersed argillaceous matter, but more importantly because organisms such as fungi and bacteria which live in the soil are capable of attacking pathogenic bacteria, and of reacting with certain other harmful substances.

As the water infiltrates deeper through the zone of aeration, the effectiveness of filtering action depends on the size of the pores, the proportion of argillaceous matter, and the distance of travel. Unconfined ground water at shallow depth is highly susceptible to pollution, but deeper ground-water bodies are recharged only by water which has been wholly or already partially purified.

In rocks with wide fissures or large openings, water infiltrates rapidly, and may join the ground-water body while still in its original polluted condition. Pollutants then mix and move along with the ground water, the degree of dilution increasing downstream until pollution becomes negligible.

In cavernous rocks, as distinct from sands, polluted water moves through a limited number of relatively large channels. Since it moves freely and quickly, the water is capable of carrying pollutants for considerable distances.

15.4.1.7 Effects of pollution intensity and distribution

For equal volumes, pollutants spread over a large soil area are far less dangerous than pollutants concentrated at a single point, because the soil's purifying surface and the volume or rock acting as a filter are much greater, and the pollution intensity over the ground-water surface is smaller per unit of area. Concentrated pollution causes massive contamination in the ground water at one point, and possibly for some distance down the hydraulic gradient.

It is obvious that in both cases, the degree of contamination to which the ground water is subject depends on the intensity of the source of pollution.

15.4.2 *Sources of pollution*

There are numerous sources of pollution.

15.4.2.1 Waste water

Urban ground-water contamination may be caused by the discharge of waste water into streams which feed the ground water, leaks from sewers, overflows or leakages from sewage tanks, or structural failures leading to the discharge of waste water on to the soil or into a stream, with a resulting massive, localized infiltration of waste water. Other causes are disused wells, leaking cesspits and septic tanks. Not only do Cl, NO_3, NO_2, SO_4 and various cations make their appearance, but pathogenic micro-organisms as well. Such waste water may include insecticides and garden fertilizers, and also detergents which facilitate the detachment and transport of micro-organisms.

15.4.2.2 Rubbish tips

The disposal of household waste is one of the activities that environmental protection seeks to control. Waste is commonly disposed of by burial in extensive dumps, which represent a hazard because they permit massive pollution of a small area and any associated ground water.

Some substances decompose by fermentation, but more rubbish is usually dumped before the self-purification process is complete, so that downward percolating rain-water is polluted during its passage to the water table. Furthermore, some toxic substances are indestructible.

If the water is at a shallow depth, the pollution gradually extends through the ground-water body in a zone more or less elongated in the direction along which the water is moving. In time this zone may stretch a considerable distance down gradient but not far up gradient. For some distance away from the tip, reduction phenomena are predominant, especially the reduction of SO_4^{--} with the formation of S^{--} and H_2S. Iron, manganese and other metals are mobile. The water also contains abundant HCO_3, CO_3, NO_3, NO_2 and Cl. At some distance from the tip, the reduction phenomena gives way to oxidation. Iron and other metals precipitate, and there is no longer any NO_2, S^{--}, or H_2S. As the area of pollution extends, the concentration of pollutants in the water is decreased by dilution, and finally becomes negligible.

A tip containing 1,000 cubic metres of rubbish can yield 1.25 tons of K and Na, 0.8 tons of Ca and Mg, 0.7 tons of chloride, 0.19 tons of sulphate and 3.2 tons of bicarbonate.

15.4.2.3 Industries and mines

Metal refineries and chemical industries, paper mills, sugar refineries, and breweries produce large quantities of waste water which is frequently discharged into streams.

Dairies, wool-washing plants, small dye works, small distilleries, canneries and laundries may dispose of their water by infiltration. Toxic substances from dye works, chrome salts from canneries, and synthetic detergents which facilitate the propagation of bacteria may all enter the ground water.

Mining wastes and water pumped from mines are capable of polluting ground water, particularly when they contain very soluble substances like chlorides. Mine tips containing pyritic schists give rise to H_2SO_4.

Ground water can be seriously polluted by petroleum products from drilling sites, refineries, reservoirs, pipelines, underground storage tanks, petrol pumps, or by accidental spillage from road-tankers, as the hydrocarbons accumulate in the upper part of the zone of saturation.

Pollution may last for a long time, because the threshold of detection by smell and taste is extremely low, being 10^{-6} for petrol and 10^{-9} for fuel oil.

It is reckoned that only 1 litre of petrol can contaminate 1,000 to 5,000 m³ of water.

15.4.2.4 Agriculture

Stock breeding can cause pollution, particularly where there is a concentration of sources of pollution such as manure heaps, liquid manure pits, and livestock pens.

Applications of fertilizers to cultivated fields are less dangerous because pollutants are not so concentrated. Nevertheless they add extra chlorides, including potassium chloride, to the ground water, as well as sulphates, nitrates, ammonium salts and calcium and ammonium phosphates.

With an input of say 80 kg of NO_3 per ha from KNO_3 and from 60 to 200 kg of Cl per ha originating from KCl, and a ground-water recharge of 200 mm of rainwater annually, there would be the not insignificant addition of 30 - 100 mg/l of Cl and 40 mg/l of NO_3. Part of this addition in fact would be absorbed by plants, but the rest is diluted with the ground water. Insecticides (1 to 2 kg per ha), herbicides (1 to 7 kg,) and fungicides (1 to 4 kg) are other forms of pollutants.

Manure does not usually cause appreciable pollution, because it is sparsely spread and its elements are absorbed by the vegetation.

Possible sources of pollution include drainage canals from irrigated areas, and irrigation water which evaporates after being applied to the land surface. This is especially likely to cause damaging accumulation of salt in the soil in arid regions.

15.4.2.5 Intensive use of ground water

Over-pumping may deplete a ground-water body sufficiently to cause an inflow of adjacent water that may be either saline or polluted. The same thing may happen if the head in the ground water body is reduced by pumping to less than that of the neighbouring ground-water bodies directly or indirectly communicating with it. This is most likely to happen when fresh ground water is withdrawn from a point near to its interface with sea water.

Other possibilities include bed and bank leakage of water from rivers, ponds or lakes.

15.4.2.6 Contamination by radioactive substances

Radioactive tracers are occasionally injected into the ground water to aid in the study of the movement of the water or of the pollutants. They should be used for this purpose only under conditions of the strictest control.

15.5 Water quality standards

15.5.1 *Drinking water standards*

Acceptable standards for drinking water may vary considerably from one country or even from one district to another. European standards laid down by the World Health Organization (WHO) in 1971, and the international standards laid down in 1972, by the same organization, are reproduced in Table 15.5.1a.

TABLE 15.5.1a. Standards for drinking water

	European standards	International standards
Biology[1]		
Coliform bacteria	Nil	Nil
Escherichia coli	Nil	Nil
Streptococcus faecalis	Nil	Nil
Clostridium perfringens	Nil	Nil
Virus	Less than 1 plaque forming unit per litre per examination in 10 litres of water	
Microscopic organisms	Nil	
Radioactivity		
Overall α radioactivity	< 3 pCi/l	< 3 pCi/l
Overall β radioactivity	< 30 pCi/l	< 30 pCi/l
Chemical elements	*(mg/l)*	*(mg/l)*
Pb	< 0.1	< 0.1
As	< 0.05	< 0.05
Se	< 0.01	< 0.01
Hexavalent Cr	< 0.05	< 0.05
Cd	< 0.01	< 0.01
Cyanides (in CN)	< 0.05	< 0.05
Ba	< 1.00	< 1.00
Cyclic aromatic hydrocarbon	< 0.20	
Total Hg	< 0.01	< 0.01
Phenol compounds (in phenol)	< 0.001	< 0.001-0.002
NO_3 recommended	< 50	
acceptable	50-100	
not recommended	> 100	
Cu	< 0.05	0.05-1.5
Total Fe	< 0.1	0.10-1.0
Mn	< 0.05	0.10-0.5
Zn	< 5	5.00-15
Mg if $SO_4 > 250$ mg l	< 30	< 30
if $SO_4 < 250$ mg l	< 125	< 125
SO_4	< 250	250-400
H_2S	0.05	

1. No 100 ml sample to contain *E. Coli* or more than 10 coliform.

	European standards	International standards
Cl recommended	< 200	
acceptable	< 600	
NH_4	< 0.05	
Total hardness	2-10 meq/l	2-100 meq/l
Ca	75-200 mg/l	75-200 mg/l

F In the case of fluorine the limits depend upon air temperature:

Mean annual maximum day time temperature (° C)	*Lower limit (mg/l)*	*Optimum (mg/l)*	*Upper limit (mg/l)*	*Unsuitable (mg/l)*
10–12	0.9	1.2	1.7	2.4
12.1–14.6	0.8	1.1	1.5	2.2
14.7–17.6	0.8	1.0	1.3	2.0
17.7–21.4	0.7	0.9	1.2	1.8
21.5–26.2	0.7	0.8	1.0	1.6
26.3–32.6	0.6	0.7	0.8	1.4

Permissible and mandatory limits of trace elements present in drinking water are discussed in Tables 15.5.1b, c, d and e.

TABLE 15.5.1b. Trace elements and compounds in drinking water (modified from Taylor, 1971)

	Comment	Recommended limit (mg/l)	Mandatory limit (mg/l)	Unofficial limit (mg/l)
Alkyl benzine Sulphonate (ABS)		0.5		
Arsenic (As)	Serious cumulative systemic poison. 100 mg usually causes severe poisoning	0.01	0.05	
Antimony (Sb)	Similar to As but less acute. Recommended limit 0.1 mg/l, routinely below 0.05 mg/l; over long periods below 0.01 mg/l	0.05	0.05	
Barium (Ba)	Muscule (including heart) stimulant. Fatal dose is 550–600 mg as chloride		1	
Beryllium (Be)	Poisonous in some of its salts in occupational exposure			None
Bismuth (Bi)	A heavy mineral in the arsenic family—avoid in water supplies			None
Boron (B)	Ingestion of large amounts can affect central nervous system	1	5	1
Cadmium (Cd)	13–15 ppm in food has caused illness		0.01	
Carbon chloroform extract (CCE)	At limit stated, organics in water are not considered a health hazard	0.200		
Chloride (Cl^-)	Limit set for taste reasons	250		

	Comment	Recommended limit (mg/l)	Mandatory limit (mg/l)	Unofficial limit (mg/l)
Chromium (hexavalent)	Limit provides a safety factor. Carcinogenic when inhaled		0.05	
Cobalt (Co)	Beneficial in small amounts; about 7 µg/day			None
Copper (Cu)	Body needs copper at level of about 1 mg/day for adults; not a health hazard unless ingested in large amounts	1.0		
Cyanide (CN)	Rapid fatal poison, but limit set provides safety factor of about 100	0.01	0.20	
Fluoride (F^-)	Beneficial in small amounts; above 2,250 mg dose can cause death	0.7–1.2	1.4–2.4	
Iron (Fe)		0.3		
Lead (Pb)	Serious cumulative body poison.		0.05	
Manganese (Mn)		0.05		
Mercury (Hg)	Continued ingestion or large amounts can damage brain and central nervous system			0.005
Molybdenum (Mo)	Necessary for plants and ruminants. Excessive intakes may be toxic to higher animals; acute or chronic effects not well known			None
Nickel (Ni)	May cause dermatitis in sensitive people; doses of 30–73 mg of $NiSO_4.6H_2O$ have produced topic effects			None
Nitrate (NO_3)	Excessive amounts can cause methemoglobinemia (blue baby) in infants	45		
Radium (Ra-226)	A bone-seeking, internal alpha emitter that can destroy bone marrow	3 pc/l		
Selenium (Se)	Toxic to both humans and animals in large amounts. Small amounts may be beneficial		0.01	
Silver (Ag)	Can produce irreversible, adverse cosmetic changes		0.05	
Sodium (Na)	A beneficial and needed body element, but can be harmful to people with certain diseases			200
Strontium-90	A bone-seeking internal beta emitter	10 pc/l		
Sulphate (SO_4)		250		
Zinc (Zn)	Beneficial in that a child needs 0.3 mg/kg/day; 675–2,280 mg/l may be an emetic		5	

TABLE 15.5.1c. Pesticides and herbicides found in drinking water

Pesticides[1]	Limit (mg/l)	Herbicides[2]	Limit (mg/l)
Aldrin	0.017	Algicide	1.0
Chlordane	0.003	Copper	
DDT	0.042	key (El)	1.0 (Cu)
Dieldrin	0.017	Cuprose	1.0 (Cu)
Endrin	0.001	Cutrine	1.0 (Cu)
Heptachlor	0.018	Diquot	0.2
Heptachlor	0.018	Dalapon	0.1
epoxide		2,4-D	1.0
Lindane	0.056	2,4,5-T	(Should not be used)
Methoxychlor	0.035		
Toxaphene	0.005		
Organic		Micro-gard (PR)	0
phosphorus		Mogul algicide	1.0
plus carbamates (such as		(AG-470)	
Parathion, Malathion,		Radapon	0.1
Carbaryl, etc.)	—[3]	Silver (2,4,5-TP)	0.2
		Tordon 1 OK and 101	0
		Twink	0.2
		Ureabor	1.0 (as B)

1. One or all of these complex organic compounds have severe, acute, adverse health effects when ingested in large amounts. Small amounts accumulate and long-term effects are generally unknown.
2. A group having toxic properties of a generally lower order than the 'Pesticides' group. However, they should be used with great care and should not be found in drinking water.
3. The organic phosphorous and carbamate pesticides are severe, acute poisons affecting the central nervous system; ingestion of small amounts over time can harm the central nervous system.

TABLE 15.5.1d. Maximum permissible isotope concentration in drinking water (mg/l) (Collins, 1960)

Isotope	Period (days)	mg/l	Isotope	Period (days)	mg/l
^{18}F	0.078	2.2×10^{-10}	^{64}Cu	0.54	1.3×10^{-10}
^{24}Na	0.62	9.1×10^{-11}	^{65}Zn	250	2.5×10^{-8}
^{32}P	14.3	6.9×10^{-11}	^{89}Sr	53	2.5×10^{-11}
^{36}Cl	1.6×10^{8}	1.7×10^{-2}	^{90}Sr	9.1×10^{3}	5.0×10^{-10}
^{45}Ca	152	5.3×10^{-10}	^{131}I	8	4.6×10^{-11}
^{51}Cr	26.5	2.1×10^{-8}	^{198}Au	2.69	2.4×10^{-10}
^{86}Mn	0.108	1.4×10^{-11}	^{210}Pb	9.1×10^{3}	2.9×10^{-10}
^{55}Fe	1.06×10^{3}	2.3×10^{-7}	W (natural)	1.64×10^{12}	0.3
^{59}Fe	43.3	2.1×10^{-10}			
^{60}Ca	1.9×10^{3}	3.6×10^{-8}			

TABLE 15.5.1c. Maximum radiation limits permissible in drinking water (µCi/l) (after IAEA, 1969)

Isotope	Limit	Isotope	Limit
^{24}Na	0.2	^{76}As	0.02
^{32}P	0.02	^{90}Sr	4×10^{-4}
^{51}Cr	1.6	^{122}Sb	0.03
^{60}Co	0.05	^{131}I	2×10^{-3}
^{64}Cu	0.3	^{137}Cs	1.5×10^{-2}
^{65}Zn	0.1		

These values cannot be universally applied, at least in so far as the major anions and cations, SO_4^{--}, Cl^-, HCO_3^-, Ca^{++}, Mg^{++}, Na^+ and the total dissolved salts are concerned.

Acceptable upper limits will depend on the climate, the type of food, the work being done by the user, and also by the salinity of the best water available. In two cases recorded in South Australia, families lived for several months on water having total salinities in excess of 5,000 mg/l, no other water being available (Ward, 1946).

In fact, two variables must be taken into account in deciding on an acceptable standard of salinity. One is the physiological tolerance of the user, and the other is the saline content of the water that is available. Even if this is of poor quality when judged by standards elsewhere the salinity may be below the upper limits to which the human body can adjust. Table 15.5.1f, based on the taste and physiological properties of water, has been suggested for use in arid regions (Schoeller, 1937, 1955).

TABLE 15.5.1f. Standards for drinking water in arid regions

	Suitability for permanent supply				Suitability for temporary supply (temporary suitability for drinking)
	Good	Fair	Moderate	Poor	
Colour	Colourless	Colourless			
Turbidity	Clear	Clear			
Odour	Odourless	Hardly Perceptible	Slight	Slight	Unpleasant
Taste at 20°	None	Perceptible	Pronounced	Unpleasant	
Total dissolved salts (mg/l)	0–500	500–1,000	1,000–2,000	2,000–4,000	4,000–8,000
Na	0–115	115–230	230–460	460–920	920–1,840
Mg	0–30	30–60	60–120	60–120	120–240
Mg/12 + Ca/20	0–5	5–10	10–20	20–40	40–80
Cl	0–177.5	177.5–355	355–710	710–1,420	1,420–2,840
SO_4	0–144	148–288	288–576	576–1,152	1,152–2,304

In Table 15.5.2a it is the element with the highest value in relation to the limits which defines the suitability of the water for drinking and not the value of the total dissolved salts.

15.5.2 *Livestock water standards*

In Tables 15.5.2a and 15.5.2b will be found a certain number of standards which may be used to determine the suitability of water supplied to livestock.

TABLE 15.5.2a. Standards for livestock water (mg/l)

Livestock	Maximum total dissolved salts (TDS)		Maximum magnesium
	Western Australia[1]	Victoria[2]	Australia[2]
Poultry	2,900	3,500	
Pigs	4,300	4,500	
Horses	6,400	6,000	250
Dairy cows	7,100	6,000	250
Beef cattle	10,000	7,000	400
Sheep on dry feed	13,000	14,000	500
Ewes with lambs	10,000	4,500	250
Ewes in milk	10,000	6,000	

1. After Western Australia Department of Agriculture, 1962.
2. After Australian Department of National Development, 1969.

TABLE 15.5.2b. Key water quality criteria for livestock use

Substance	Upper limit	Substance	Upper limit
Total dissolved salts (TDS)		*Hazardous trace elements*	
(mg/l)	10,000[1]	(mg/l)	
		As	0.05
Radionuclides (pCi/l)		Cd	0.01
^{90}Sc	10	Cr	0.05
^{225}Ra	3	F	2.40
Activity	1,000	Pb	0.05
		Se	0.01
Chemical elements (mg/l)			
$MgSO_4$	2,050		
NaCl	2,000	*Organic substances*	
SO_4	1,000	Algae (water bloom)	—[2]

1. Depending upon animal species and ionic composition of the water.
2. Avoid abnormally heavy growth of blue-green algae. Parasites and pathogens conform to epidemiological evidence. Disolved organic substances.

Factors affecting salt tolerance in livestock. The maximum concentrations of salts in drinking water may be higher or lower than the limits indicated, depending upon circumstances.

Greater absorption of water in summer-time or in an arid climate, or when dry feed is being given, should be accompanied by a lowering of the limit.

Most animals tolerate NaCl better than $CaCl_2$ or $MgCl_2$.

15.5.3 *Irrigation water standards*

The salinity of water suitable for irrigation not only depends on the chemical composition but also on the nature of the plants to be irrigated, the soil type, the climate, the method of irrigation, and drainage.

Chemical composition. In judging the suitability of the water for irrigation, use is made of the relationship between per cent sodium:

$$\frac{(r\mathrm{Na} + r\mathrm{K}) \times 100}{r\mathrm{Ca} + r\mathrm{Mg} + r\mathrm{Na} + r\mathrm{K}}$$

and electrical conductivity (EC) or the total dissolved salts, as given in the diagram by Wilcox (Todd, 1959) reproduced in Figure 15.5.3a.

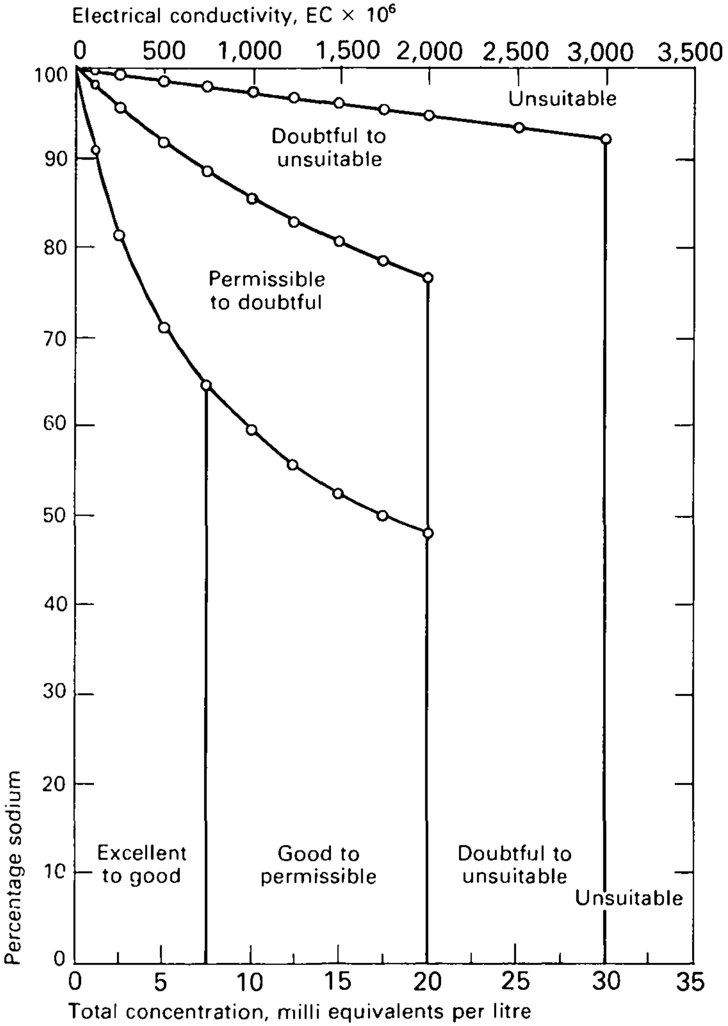

FIG. 15.5.3a. Diagram for irrigation water classification based on electrical conductivity and per cent sodium (after Wilcox in Todd, 1959).

The accepted formula is $EC \approx (r \text{ anions} + r \text{ cations})\ 100 = \frac{\text{TDS mg}}{0.7}$.

Use may also be made of the relationship between the sodium absortium ratio (SAR) and the electrical conductivity (EC) of the water as in Figure 15.5.3b.

It will be recalled that

$$\text{SAR} = \frac{r\text{Na}}{\left(\frac{r\text{Ca} + r\text{Mg}}{2}\right)^{1/2}}$$

The United States Salinity Laboratory has devised a system for classifying irrigation water. This system is reproduced in Table 15.5.3a.

TABLE 15.5.3a. United States Salinity Laboratory classification of irrigation waters

Total dissolved salts (TDS) (mg/l)	EC × 10^6 (μmhos/cm at 25° C)	Class	Remarks
< 200	< 250	C_1	Low salinity water—can be used for irrigation with most crops on most soils with little likelihood that a salinity problem will develop. Some leaching is required, but this occurs under normal irrigation practices except in soils of extremely low permeability.
200–500	250–750	C_2	Medium salinity water—can be used if a moderate amount of leaching occurs. Plants with moderate salt tolerance can be grown in most instances without special practices for salinity control.
500–1,500	750-2,250	C_3	High salinity water—cannot be used on soils with restricted drainage, special management for salinity control may be required and plants with good salt tolerance should be selected.
1,500-3,000	2,250–5,000	C_4	Very high salinity water—is not suitable for irrigation under ordinary conditions but may be used occasionally under very special circumstances. The soil must be permeable; drainage must be adequate; irrigation water must be applied in excess to provide considerable leaching and very salt tolerant crops should be selected.

A complementary classification given in Table 15.5.3b has been established on the basis of the sodium content of the water.

Irrigation water quality standards adopted in the U.S.S.R. are listed in Table 15.5.3c.

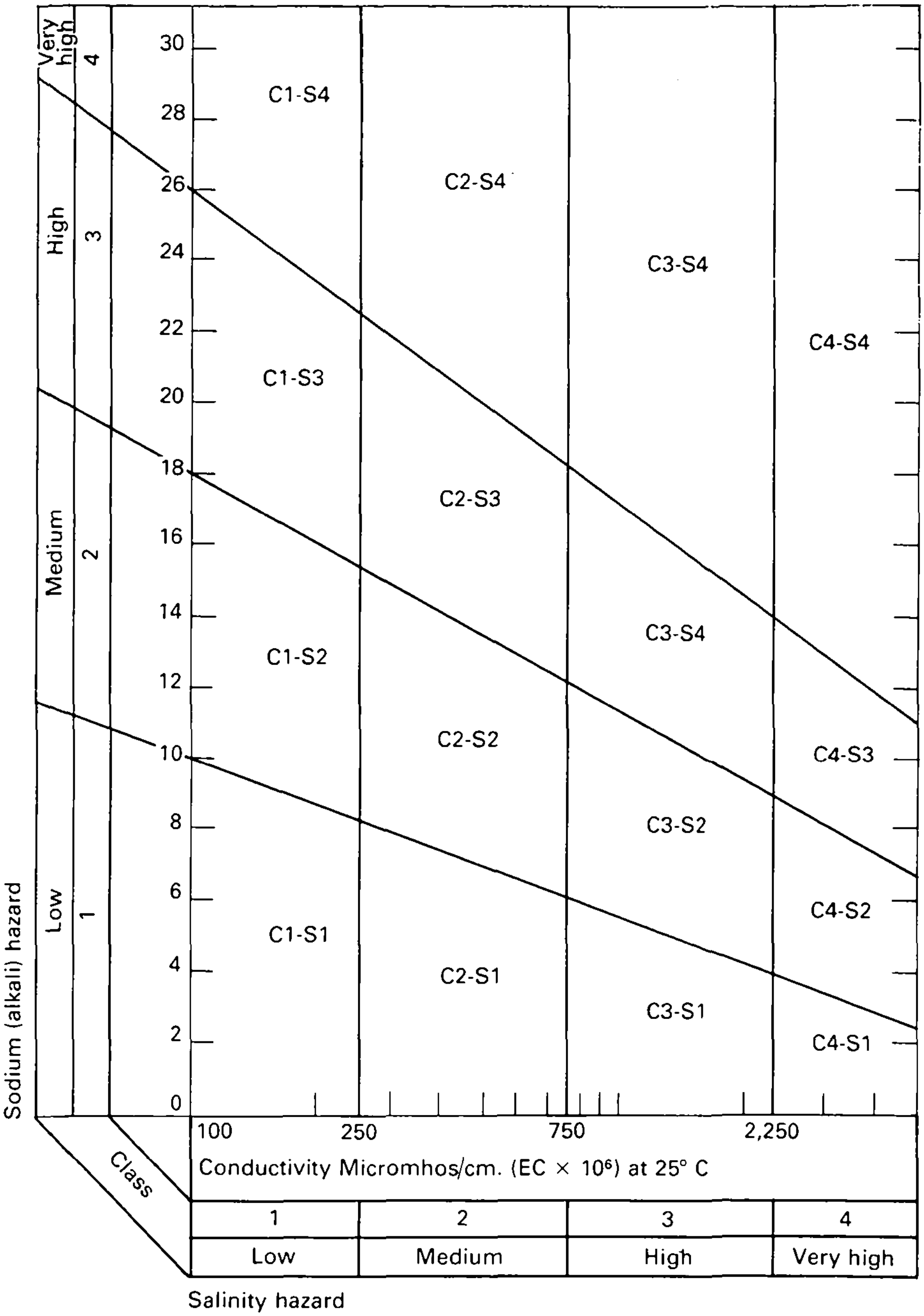

FIG. 15.5.3b. Quality criteria for irrigation water.
(*Source*: United States Department of Agriculture.)

TABLE 15.5.3b. Irrigation water classification based on sodium content

Classification	Comment
S_1	Low sodium water can be used for irrigation on almost all soils with little danger of the development of harmful levels of exchangeable sodium. However, sodium sensitive crops such as stonefruit trees and avocado may accumulate injurious concentrations of sodium.
S_2	Medium sodium water will present an appreciable sodium hazard in fine textured soils having high cation exchange capacity, especially under low leaching conditions, unless gypsum is present in the soil. This water may be used on coarse-textured or organic soils with good permeability.
S_3	High sodium water may produce harmful levels of exchangeable sodium in most soils and will require special soil management—good drainage, high leaching and organic matter additions. Gypsiferous soils may not develop harmful levels of exchangeable sodium from such waters. Chemical amendments may be required for replacement of exchangeable sodium, except that amendments may not be feasible with waters of very high salinity.
S_4	Very high sodium water is generally unsatisfactory for irrigation purposes except at low and perhaps medium salinity, where the dissolving of calcium from the soil, or the use of gypsum or other additives may make the use of these waters feasible.

TABLE 15.5.3c. U.S.S.R. classification of irrigation water

Total dissolved salts (TDS) (mg/l)	Comment
200–500	Best quality water
1,000–2,000	Water involving a hazard
3,000–7,000	Water which can be used for irrigation only with leaching and perfect drainage

In Table 15.5.3d crops are classified according to their relative tolerance to salt concentrations present in soil moisture, in the light of experience gained in the United States.

TABLE 15.5.3d. Relative tolerance of crops to salt concentrations (United States Department of Agriculture, 1954)

Crop division	Low salt tolerance	Medium salt tolerance	High salt tolerance
Fruit crops	Avocado Lemon Strawberry Peach Apricot Almond Plum Prune Grapefruit Orange Apple Pear	Cantaloupe Date Olive Fig Pomegranate	Date palm

Crop division	Low salt tolerance	Medium salt tolerance	High salt tolerance
Vegetable crops[1]	$EC \times 10^3 = 3$ Green beans Celery Radish $EC \times 10^3 = 4$	$EC \times 10^3 = 4$ Cucumber Squash Peas Onion Carrot Potatoes Sweet corn Lettuce Cauliflower Bell pepper Cabbage Broccoli Tomato $EC \times 10^3 = 10$	$EC \times 10^3 = 10$ Spinach Asparagus Kale Garden beet $EC \times 10^3 = 12$
Forage crops[1]	$EC \times 10^3 = 2$ Burnet Ladino clover Red clover Alsike clover Meadow foxtail White dutch clover $EC \times 10^3 = 4$	$EC \times 10^3 = 4$ Sickle milkvetch Sour clover Cicer milkvetch Tall meadow oatgrass Smooth brome Big trefoil Reed canary Meadow fescue Blue grama Orchard grass Oats (hay) Wheat (hay) Rye (hay) Tall fescue Alfalfa Hubam clover Sudan grass Dallis grass Strawberry clover Mountain brome Perennial rye grass Yellow sweet clover White sweet clover $EC \times 10^3 = 12$	$EC \times 10^3 = 12$ Bird's-foot trefoil Barley (hay) Western wheat grass Canada wild rye Rescue grass Rhodes grass Bermuda grass Nuttall alkali grass Salt grass Alkali sacaton $EC \times 10^3 = 18$
Field crops[1]	$EC \times 10^3 = 4$ Field beans	$EC \times 10^3 = 6$ Castorbeans Sunflower Flax Corn (field) Sorghum (grain) Rice Oats (grain) Wheat (grain) Rye (grain) $EC \times 10^3 = 10$	$EC \times 10^3 = 10$ Cotton Rape Sugar beet Barley (grain) $EC \times 10^3 = 16$

1. Electrical conductivity (EC) values represent salinity levels at which a 50 per cent decrease in yield may be expected as compared to yield on non-saline soil under comparable growing conditions. Concentrations refer to soil water.

Boron is essential to plant growth, but only in very small quantities. It is very toxic to some plants, in concentrations that would not affect others that are more tolerant, as shown in Tables 15.5.3e and f.

TABLE 15.5.3e. Permissible limits of boron in irrigation waters (mg/l)

Class of water	Sensitive crops	Semi-tolerant crops	Tolerant crops
Excellent	< 0.33	< 0.67	< 1.00
Good	0.33–0.67	0.67–1.33	1.00–2.00
Permissible	0.67–1.00	1.33–2.00	2.00–3.00
Doubtful	1.00–1.25	2.00–2.50	3.00–3.75
Unsuitable	> 1.25	> 2.50	> 3.75

TABLE 15.5.3f. Relative tolerance of plants to boron

Sensitive	Semi-tolerant	Tolerant
Lemon	Lima bean	Carrot
Grapefruit	Sweet potato	Lettuce
Avocado	Bell pepper	Cabbage
Orange	Pumpkin	Turnip
Apricot	Zinnia	Onion
Peach	Oat	Broadbean
Cherry	Milo	Gladiolus
Persimmon	Corn	Alfalfa
Kadota fig	Wheat	Gardenbeet
Grape	Barley	Mangel
Apple	Olive	Sugarbeet
Pear	Field Pea	Date Palm
Plum	Radish	Asparagus
Navy bean	Sweetpea	Athel (*Tamarix aphylla*)
Jerusalem artichoke	Tomato	
Walnut	Cotton	
Pecan	Potato	
	Sunflower	

In each group, the first named plants are considered the more sensitive, the last named as more tolerant.

Concentrations of other trace elements greater than those indicated in Table 15.5.3g may also adversely affect plant growth.

TABLE 15.5.3g. Limits of trace elements in irrigation waters

Element	Concentration (mg/l)	Element	Concentration (mg/l)
Beryllium	0.5	Copper	0.2–5
Cadmium	0.005	Zinc	10
Chromium	5		

The nature of the soil has an influence on plant tolerance. The tolerance is in relation to the soil conductivity, less for clay and loam than for sand. The following values are the ratios of the extreme conductivities compared with the sandy soil conductivity: sandy soil, 1; loamy soil, 0.6; loam, 0.45–0.6; loamy clay, 0.30; clay, 0.16. Soils having lower comparative conductivity are less tolerant to salt.

15.5.4 *Quality of water for industry*

It is not possible in this guide to discuss in any detail the quality requirements of industrial water since these differ widely from one industry to another. Reference should be made to the *Report of the National Technical Advisory Committee to the Secretary of the Interior Water Quality Criteria* (1968).

Walton (1970) quotes the recommendations of the committee on quality tolerances of water for industrial use concerning water suitable for boiler feed. These are reproduced in Table 15.5.4a.

TABLE 15.5.4a. Tolerance limits for boiler-feed water[1] (mg/l)

	Pressure (psi)[2]			
	0-150	150-250	250-400	Over 400
Turbidity	20	10	5	1
Colour	80	40	5	2
Oxygen consumed	15	10	4	3
Dissolved oxygen	1.4	0.14	0	0
Hydrogen sulphide (H_2S)	5	3	0	0
Total hardness as $CaCO_3$	80	40	10	2
Sulphate carbonate ratio Na_2SO_4/Na_2CO_3	1/1	2/1	3/1	3/1
Aluminium oxide (Al_2O_3)	5	0.5	0.05	0.01
Silica (SiO_2)	40	20	5	1
Bicarbonate (HCO_3)	50	30	5	0
Carbonate (CO_3)	200	100	40	20
Hydroxide (OH)	50	40	30	15
Total solids	3,000–500	2,500–500	1,500–100	50
pH value (minimum)	8.0	8.4	9.0	9.6

1. It is also required that there be no corrosiveness or slime formation.
2. 1 psi = 0.07031 kg/cm².

From an extensive study of industrial water quality, a committee of New England Water Works Association (Anon, 1940) suggested limits for selected industries given in Table 15.5.4b.

TABLE 15.5.4b. Suggested water-quality tolerances for industrial uses (allowable limits in mg/l)[1]

Industry or use	Turbidity	Colour	Odour and taste	Iron as Fe	Manganese as Mn	Total solids	Hardness as $CaCO_3$	Alkalinity as $CaCO_3$	Hydrogen sulphide	Health	pH	Other requirements
Air conditioning	...	...	Low	0.5	0.5	...	...	...	1.0	...		No corrosiveness or slime formation
Baking	10	10	Low	0.2	0.2	...	...	...	0.2	Potable	...	
Brewing and distilling												NaCl 275
Light beer, gin	10	...	Low	0.1	0.1	500	...	75	0.2	Potable	6.5–7.0	
Dark beer, whiskey	10	...	Low	0.1	0.1	1,000	...	150	0.2	Potable	7.0	
Canning												
Legumes	10	...	Low	0.2	0.2	...	25–75	...	1.0	Potable	...	
General	10	...	Low	0.2	0.2	...	...		1.0	Potable	...	
Carbonated beverages	2	10	Low	0.2	0.2	850	250	50–100	0.2	Potable	...	Organic matter infinitesimal, oxygen consumed 1.5
Confectionery	...	...	Low	0.2	0.2	100	...	...	0.2	Potable	7.0	
Cooling	50	...	...	0.5	0.5	...	50	...	5	...	...	No corrosiveness or slime formation
Food, general	10	...	Low	0.2	0.2	...	...	...	...	Potable	...	
Ice	5	5	Low	0.2	0.2	1,300	...	...	...	Potable	...	SiO_2 10
Laundering	...	...	...	0.2	0.2	...	50	...	...	...	...	
Plastics, clear	2	2	...	0.02	0.02	200	...	...	...	...		
Paper and pulp												No grit or corrosiveness
Ground wood	50	20	...	1.0	0.5	...	180	...	...	...	...	
Kraft pulp	25	15	...	0.2	0.1	300	100	...	...	...	...	
Soda and sulphide pulp	15	10	...	0.1	0.05	200	100	...	...	...	...	
High-grade light papers	5	5	...	0.0	0.05	200	50	...	...	...	...	No slime formation

Rayon (viscose)												
Pulp production	5	5	...	0.05	0.03	100	8	50	...	...	...	OH 8, Al_2O_3 8, SiO_2 25, Cu 5
Manufacture	0.3	...	...	0.0	0.0	...	55	...	...	...	7.8–8.3	
Steel manufacture	...	...	...	...	...	...	50	...	...	...	6.8–7.0	Temperature 75° F, Cl 175, suspended matter 25, minimum organic content and corrosiveness
Sugar manufacture	...	...	...	0.1	...	...	...	...	...	...	...	Ca 20, Mg 10, SO_4 20, Cl 20, HCO_3 100
Synthetic rubber	...	...	...	...	...	...	50	...	...	...	...	Oxygen consumed 3.0, minimum organic content and corrosiveness
Tanning	20	10–100	...	0.2	0.2	...	50–135	135	...	...	...	OH 8
Textiles												
General	5	20	...	0.25	0.25	...	...	...	...	...	...	Constant composition,
Dyeing	5	5–20	...	0.25	0.25	200	...	...	...	...	...	residual alumina < 0.5

1. Limits concerning boiler-feed water originally given by the author have been deleted since they appear in Table 15.5.4a.

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Appendixes

A Selected general references in ground-water hydrology

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B List of symbols used in text

A	Change in moisture storage during rise in water table; areal extent of water-balance area; constant.
a	Hydraulic diffusivity; coefficient of piezoconductivity (U.S.S.R.); heat conduction in rocks; constant.
a_y	Coefficient of water-level conductivity (U.S.S.R.).
B	Moisture depletion during fall in water table; increase in moisture stored in zone of aeration between land surface and minimum position of water table; heat exchange with soil; constant.
B_b	Barometric efficiency.
b	Regression coefficient; constant.
C	Coefficient of variation; constant concentration of CO_2 in the atmosphere, molal concentration.
C_a	Absorption capacity.
C_{ws}	Specific heat of water/solid mixture.
C_1, C_2	Moisture reserves in the zone of aeration between the land surface and water table at the beginning and end of a computational period.
C_{l_n}	Chloride concentration in ground water.
C_{l_n}	Chloride concentration in rainwater.
c_{pw}	Specific heat of water.
D	Diffusivity coefficient.
d	Absolute deviation from arithmetical mean; drain width; distance between two geophones; thickness of zone of aeration.
d_e	Effective grain diameter.
$d_{1,c}$	Change in quantity of salts at land surface.
$d_{2,c}$	Change in water-soluble salts in the zone bounded by the land surface and the highest position of capillary fringe.
$d_{3,c}$	Change in water-soluble salts within the capillary fringe.
E	Quantity of water leaving or entering confined aquifer.
EC	Electrical conductivity log.
E_s	Bulk modulus of elasticity of aquifer.
E_w	Bulk modulus of elasticity of water.
E_x	Component of the telluric field for the direction O_x.
F	Area; cross-sectional area; flowmeter log; formation factor of an aquifer.
f, f_m	Function of. . . .
f_k	Amount of infiltrating water from irrigation canals.
f_n	Components of recharge to unconfined ground water from irrigation water.
GR	Gamma-ray log.
GGR	Gamma-gamma-ray log.
g	Acceleration due to gravity.
g_0	Initial moisture content.
H	Rise or fall in water table; saturated thickness of unconfined aquifer; depth of seismic investigations.
H, H_{max}, H_{min}	Hydraulic head in prototype.
H_{av}	Average saturated thickness of unconfined aquifer.
H_e	Height of capillary rise.
H_s	Depth of water over ground surface.
H_1, H_2	Ground-water level.

H_i, H_m	Ground-water level; head.
Hy	Component of the magnetic field for the direction Oy.
h	Saturated thickness of unconfined aquifer.
h_i'	Thickness of layer *i*.
h_m	Moisture storage in the zone of aeration.
I	Hydraulic gradient; recharge; infiltration to ground-water reservoir.
i	Intensity of precipitation; electrical intensity.
K	Caliper log; hydraulic conductivity; geometric coefficient for resistivity measurements; temperature coefficient for spontaneous potential.
K_R	Coefficient of surface run-off.
K_b	Vertical hydraulic conductivity of semi-confining layer.
K_f	Hydraulic conductivity along fractures perpendicular to the bedding.
K_k	Coefficient of turbulent flow.
K_t	Effective hydraulic conductivity.
K_n	Coefficient of ground-water discharge.
K_r	Horizontal hydraulic conductivity; hydraulic conductivity to radial flow.
K_z	Hydraulic conductivity parallel to axis of well; vertical hydraulic conductivity.
K_1	Water-vapour condensation at land surface and in zone of aeration.
K_1	Water condensation only in zone of aeration.
k	Intrinsic permeability; water conductivity coefficient; negative recession constant.
k_t	Thermal conductivity.
k_{ws}	Thermal conductivity of water/solid mixture.
k_1, k_2	Constant.
L	Dimensions of length; distance between observation points in an aquifer.
LZ	Heat losses by evaporation.
L_w	Distance of wetting front from land surface in zone of aeration.
l	Width of fractures.
l_1, l_2	Distance between observation wells.
M	Dimensions of mass; pore-water pressure; mineralization of water.
M_n	Annual mean modulus of ground-water discharge to river; coefficient of base flow.
M_{ni}	Modulus of ground-water discharge from single aquifer to river.
m	Positive integer; true arithmetical mean; thickness of aquifer; mass.
m_b	Thickness of semi-confining layer.
m_p	Scaled thickness of semi-confining layer.
N	Precipitation; neutron-gamma-ray log; maximum soil-moisture capacity.
NV	Volume moisture content after gravity drainage.
N_R	Reynolds number.
N_1, N_2	Surface areas.
n	Number of samples; exponent, pore distribution coefficient.
n_c	Amount of salts introduced with precipitation.
n_1, n_2	Salts introduced to or lost from atmosphere respectively in the form of solid or liquid particles.
P	Precipitation; heat exchange with atmosphere; pressure.
PV	Maximum moisture capacity.
P_0	Amplitude of temperature fluctuations at land surface.
P_x	Amplitude of temperature fluctuations at depth *x*.
Q	Discharge.
Q_{asc}	Volume of ascending ground water.

Q_b	Rate of unconfined ground-water leakage into or out of underlying semi-confined aquifer.
$Q_{b,1}$	Confined leakage from underlying aquifer.
$Q_{b,2}$	Unconfined ground-water outflow downward as leakage to underlying semi-confined aquifer.
Q_{gw}	Annual ground-water discharge to river.
Q_n	Rate of ground-water discharge to river.
Q_{nc}	Rate of artesian overflow.
Q_0	Initial discharge.
Q_t	Discharge after time t.
Q_w	Rate of discharge.
Q_1, Q_2	Rates of ground-water inflow into and outflow from element of flow respectively.
Q_I	Ground-water inflow into basin.
Q_{II}	Ground-water outflow from basin.
Q''_{II}, Q'_{II}	Ground-water outflow from above and below moraine respectively.
q	Minimum monthly base flow; specific discharge.
$q_{a,c}$	Amount of salts removed from the land surface by infiltration into the zone of aeration.
$q_{w,c}$	Quantity of water-soluble salts, introduced by infiltration, flowing out of subzone.
$q'_{w,c}$	Quantity of water-soluble salts moving from capillary zone into zone of saturation.
$q_{1,c}$ $q_{2,c}$	Quantity of water-soluble salts entering or leaving elemental column of unconfined ground water respectively.
R	Infiltration rate; radiation balance; multiple correlation coefficient.
RES	Electrical resistivity log.
R_h	Hydraulic radius.
R_r, R_w	Electrical resistance.
$R(\lambda)$	Function (λ).
r	Correlation coefficient; temperature fluctuation period; radial distance; radius.
r_e	Extent of cone of depression around discharging well at equilibrium.
r_w	Effective radius of discharging well.
r_1, r_2	Distance to observation wells from discharging well.
S	Coefficient of storage (S' = apparent coefficient of storage); specific yield; sonic log; standard deviation.
SP	Spontaneous (or self-) potential log.
S_c	Change in liquid content across a wetting front.
S_e	Standard error of estimate.
S_p	Standard error of prediction.
S_x	Standard error of mean.
S_y	Standard deviation of the values of the dependent variables.
$S_{y \cdot x}$	Standard error of Y.
S^2X	Variance of X.
S^a, S^r	Salt storage in the zone of aeration and ground-water reservoir respectively.
S_1^a, S_2^a	Initial and final amounts of salts respectively stored in the zone of aeration.
S_3^a	Salts introduced with atmospheric precipitation (= S_3^0).
S_4^a	Salts introduced with irrigation water (= S_4^0),
S_5^a	Salts introduced by processes of cultivation (= S_6^0).

S_6^a	Salts introduced with fertilizers.
S_7^a	Salts moving upward from zone of saturation by processes of evaporation, transpiration, diffusion, etc. ($= S_6^r$).
S_8^a	Salts lost by downward percolation of precipitation and irrigation water into unconfined ground-water reservoir ($= S_3^r + S_4^r$).
S_9^a	Salts lost during crop harvesting ($= S_{10}^o$).
S_1^o, S_2^o	Initial and final amounts of salts respectively stored in surface streams and reservoirs, in the zone of aeration, and in the uppermost ground-water reservoir down to underlying impervious bed.
S_3^o	Salts introduced with atmospheric precipitation ($= S_3^a$).
S_4^o	Salts introduced with irrigation water ($= S_4^a$).
S_5^o	Salts introduced by ground-water inflow ($= S_5^r$).
S_6^o	Salts introduced by processes of cultivation ($= S_5^a$).
S_7^o	Salts introduced with fertilizers.
S_8^o	Salts removed by ground-water outflow ($= S_7^r$).
S_9^o	Salts removed by drainage water ($= S_8^r$).
S_{10}^o	Salts removed by crop harvesting ($= S_9^a$).
S_1^r, S_2^r	Initial and final amounts respectively of salts stored in the ground-water reservoir.
S_3^r	Salts introduced with percolating precipitation.
S_4^r	Salts introduced with percolating water.
S_5^r	Salts introduced by ground-water inflow ($= S_5^o$).
S_6^r	Salts transferred to the zone of aeration from ground-water reservoir by evaporation or transpiration ($= S_7^a$).
S_7^r	Salts removed by ground-water outflow ($= S_8^o$).
S_8^r	Salts removed by drainage water ($= S_9^o$).
S'	Apparent coefficient of storage.
s, s_1, s_2	Drawdown in observation well.
s_{nc}	Drawdown in overflowing well.
s_w	Drawdown in pumped well.
T	Dimensions of time; period of time; transmissivity; temperature log; air temperature.
T_1, T_2	Seismic intercepts.
T_x	Highest transmissivity in direction of x.
T_y	Highest transmissivity in direction of y.
t	Time; standard unit of deviation; Student's t test; period of the telluric current.
t_1	Time by which temperature cycle at depth z lags temperature cycle at land surface.
U, $\bar{U}_m$, $\bar{U}_w$	Electrical potential.
$\bar{U}^o$	Correction to electrical potential.
u	Discharge from ground-water reservoir by evaporation.
V	Volume of water moving per unit of time through cross-sectional area.
V_1, V_2, V_m	Velocities of seismic waves.
V_b	Rate of upward ground-water flow.
V_o	Specific retention.
v	Velocity of ground-water flow.

W	Rate of free ground-water recharge from above; infiltration rate.
W_a	Rate of infiltration at land surface (or evaporation if negative).
W_h, W_k	Initial and final values of moisture storage respectively in zone of aeration down to capillary fringe.
$W(u)$	Well function of u.
W_1	Increase in surface-water storage.
W_2	Change of moisture storage in zone of aeration.
X	Precipitation.
$\overline{X}$	Arithmetic mean.
X_o	Mean precipitation.
$X_{1-1'}$	Precipitation over catchment area.
X_ψ	Part of precipitation reaching water table.
x	Distance.
$\overline{x}$	Arithmetic mean.
Y_b	Baseflow discharge to rivers; quantity of ground water discharging at land surface at boundary with water-balance area.
Y_{gw}	Infiltration equivalent to annual ground-water discharge.
Y_n	Quantity of irrigation water.
Y_o	Mean run-off.
$Y_{1,ov}$	Overland flow into basin.
$Y_{2,ov}$	Overland flow out of basin.
Y_1	Input to irrigation area.
Y_2	Output from irrigation area.
$y_{1,c}$	Amount of salts deposited by surface run-off.
$y_{2,c}$	Amount of salts removed by surface run-off.
Z	Evaporation; evapotranspiration.
$Z_{1-1'}$	Evaporation over catchment.
z	Height above selected datum; depth of moistening; depth below land surface.

α	Empirical coefficient; decimal part of seepage loss from canal to unconfined ground-water reservoir.
α_R	Scale or horizontal resistance.
α_T	Scale factor of transmissivity.
α_{av}	Average solute concentration of ground water.
α_c, α_{nc}	Lag coefficient for observation well.
α_i	Salinity of a layer i.
α_i'	Mineralization of ground-water layer i.
α_n	Salt concentration in atmospheric precipitation.
α_1, α_2	Concentration of salts in surface-water inflow and outflow respectively on land surface.
$\alpha\mu$	Coefficient of absorption for a given ionic strength μ.
β	True value of slope of regression line.
β_w	Compressibility of water (reciprocal of bulk modulus of elasticity).
β_s	Compressibility of aquifer skeleton (rock materials).
β_1	Dimensionless parameter.
β_1, β_2	Concentration of salts in ground-water inflow and outflow respectively.
β^*	Over-all compressibility of aquifer (porous media and contained water).
γ	Specific weight; temperature; universal gravitational constant.
γ_{AO}	Average ambient temperature at surface.
γ_{la}	Activity factor of Ca.
γ_l, γ_o, γ_z	Temperature.
γ_o, γ_1, γ_2, γ_3	Absolute temperature, activity factors of CO_2; HCO_3; CO_3.

$\Delta\gamma_z$	Amplitude of sinusoidal component of temperature.
∂	Dynamic viscosity; density; partial pressure of CO_2 in the atmosphere.
Δd	Change in salt storage in top 1 to 5 cm layer of soil and on land surface.
ξ_1, ξ_2, ξ_3	Resistance values.
Δe	Absolute moisture gradient.
ΔH, ΔH_0	Water-level change; piezometric head above water surface.
ΔH_b	Reduction in head or difference between levels of unconfined and confined ground-water in vertical direction.
Δh	Depth interval; vertical change in water level.
θ	Volume moisture constant; porosity; function of. . . .
θ_i'	Porosity of layer *i*.
ΔL	Element of length.
μ	Specific yield (U.S.S.R.), ionic strength.
μ^*	Coefficient of storage (U.S.S.R.).
ν	Kinematic viscosity; coefficient of recharge.
ν'	Diffusity coefficient of an anistropic aquifer in the direction of observation well.
ρ	Resistivity.
ρ_a	Apparent resistivity.
ρ_m	Specific electrical resistance; mud resistivity.
ρ_w	Density of water; resistivity of water.
ρ_{ws}	Density of water/solid mixture.
ρ_1	Volume weight of dry earth material.
σ	True standard deviation.
ΔS	Change in water-soluble salts in element of ground-water flow.
ΔU	Potential difference.
τ	Period of temperature oscillation at the surface.
Φ	Hydraulic potential.
φ_k	Total water loss through canal seepage.
$\Delta\tau$	Period of infiltration of precipitation.
ΔT	Gradient of air temperature.
Δt	Interval of time; ambient temperature at surface.
Δt_x	Temperature at depth *x*.
Δv	Period of evaporation of unconfined ground water.
Δx	Element of length; distance between wells.
Δy	Element of length.
Δz	Element of length; depth interval.

List of definitions of terms used in text

Aquiclude. See *Confining bed.*

Aquifer. A formation, part of a formation or group of formations that will yield water to wells.

Arid zone. A territory with a dry (arid) climate where evaporation exceeds precipitation. Perennial streams are absent, except rivers which have their source outside the arid zone.

Artesian water. Ground water confined between impervious layers. When the artesian aquifer is tapped, the water level rises above the base of the confining bed. With a sufficient head artesian water can overflow at the land surface.

Average long-term stream run-off. The average value of streamflow for a long-term period including an equal number of full cycles of dry-and-wet years.

Bound water. Ground water physically or chemically bound to the solid matter of rocks. It is immovable in contrast to gravitational water. There are two groups of bound water: (a) water inside the solid matter of the rock; and (b) water in minute pores and fissures of rocks or at the surface of the rock held by molecular attraction.

The first group includes: (a) constitutional water; (b) water of crystallization; (c) hydrate or zeolite water.

The second group includes: (a) pellicular (loosely bound) water; (b) hygroscopic (strongly bound) water.

Capillary fringe. The zone separating the zone of aeration and the saturated zone. The capillary fringe is connected hydraulically with the saturated zone. In the capillary fringe capillary pores, fissures and other interstices contain water held by capillary forces.

Coefficient of water balance (after A. I. Kostyakov). Ratio of the product of the precipitation P (in mm) and the surface run-off index μ to the evaporation E (in mm) for the same period: $\mu P/E$.

Condensation of water vapour. Transfer of moisture from vapour into liquid.

Cone of depression. Lowering of water level or potentiometric surface as a result of pumping from a well, pit, etc. The maximum lowering is observed near the well, the minimum lowering away from the well.

Confined water. Water existing in aquifers located between impermeable layers. Confined water differs from free ground water in that its surface is not connected directly with the atmosphere and there is a positive hydrostatic pressure at the top of the aquifer.

Confining bed. A relatively impermeable bed confining aquifers.

Desorption. Return into solution of a sorbate. The phenomenon is the opposite of sorption.

Diffusion. A process of distribution of dissolved matter in a solution by means of convection from places of higher concentration to places of lower concentration.

Drainage. The removal of ground water from an aquifer by means of intercepting channels—natural or man-made.

Drainage divide. Boundary between drainage basins of two water reservoirs or streams. Surface and subsurface (ground-water) divides are distinguished.

Elevation head. The elevation of a point above a reference datum.

Evaporation. The process of transformation of water and ice into water vapour. The amount of water or ice transformed into vapour.

Potential evaporation. Maximum possible evapotranspiration under given meteorological conditions with an infinite supply of water at the evaporating surface.

Geothermal depth. Vertical distance in meters from the land surface to the depth at which the rock temperature rises by 1° C.

Geothermal gradient. Rate of change of temperature with depth. Usually expressed as °C per 100 m or °F per 100 ft.

Granulometric composition of rocks. Percentage by weight of fractions or different grain sizes composing a rock.

Ground-water régime. A description of the occurrence of ground water including recharge discharge, movement, temperature and chemical composition and environmental factors in respect to climate, hydrology, geology, soils, biology and man-imposed stresses.

Head gradient. The value of head loss per unit length of ground-water flow-path: $I = \Delta h/\Delta L$.

Hydraulic conductivity (permeability). Ability of rocks to transmit water. Hydraulic conductivity depends on the geometry of the rock pores and the kinematic viscosity of the water. It may be expressed in such units as cm/sec, m/sec, m/day.

Hydraulic diffusivity. The value characterizing the rate of transmission of head in an aquifer. The hydraulic diffusivity (a) is expressed in m^2/day or cm^2/sec and found from the formula:

$$a = \frac{Km}{S} = \frac{T}{S} = \frac{K}{S_s}$$

where:

K = hydraulic conductivity of the water-bearing material;
m = thickness of the aquifer;
T = transmissivity of the aquifer;
S = coefficient of storage of the aquifer;
S_s = specific storage of the aquifer.

Hydraulic gradient. Change in hydrostatic head per unit of distance in a given direction, usually in the direction of flow.

Hydrograph. A graph showing the change of water level or discharge in time.

Infiltration. Percolation of water into soil.

Insolation. Radiation received on the earth's surface from solar rays. Insolation produces heat, light and biological effects.

Karst water. Ground water in fissures, joints, channels and cavities formed mainly by leaching and solution of rocks by ground water but also by mechanical erosion of particles and rock debris.

Leaching. Transition into solution of a substance from a mineral or rock without disturbance of the crystalline lattice or rock structure, unlike solution when crystals or rocks are broken down.

Lysimeter. Apparatus for measuring the quantity of water percolating downwards through soil layers. It consists of a vessel in which a soil monolith or loose soil is placed, a vessel for collecting water percolated through the monolith or loose soil, and a measuring vessel for measuring the volume of percolated water.

Osmosis. Diffusion of matter (usually from solutions) through a semi-permeable medium separating two solutions of different concentration.

Piezometer. A device, such as a well, used to measure the hydrostatic head at a point.

Porosity. The total volume of all voids in rocks. The porosity is usually expressed as a percentage or ratio of voids to the total rock volume. Porosity may be (a) total (absolute, physical). i.e. the total volume of all voids irrespective of their forms, size and arrangement, or (b) effective or dynamic, i.e. the volume of the connected pores through which water flows; it is expressed by the ratio of the pore volume that is not occupied by rock-bound water or gas to the total rock volume.

Potential evaporation. See *Evaporation.*

Radiation balance of the atmosphere. Ratio of radiant energy absorbed to that emitted by the atmosphere.

Saturation zone. The part of the earth's crust in which all voids are typically filled with water under hydrostatic pressure.

Soil-moisture capacity. Capability of soils to hold water in pores on the surfaces of particles. Soil-moisture capacity is expressed by moisture content as a percentage of the weight of dry soil. There are several kinds of soil-moisture capacity:

Hydroscopic-moisture capacity. Maximum amount of water that is strongly bound with the surface of soil particles by molecular forces.

Specific retention (maximum molecular moisture capacity). Maximum amount of hygroscopic and pellicular water held by soil or rock particles.

Capillary-moisture capacity. Maximum amount of water held in capillary pores.

Maximum moisture capacity. Maximum amount of water held by soil or rock at full saturation.

Specific capacity. The rate of discharge from a well divided by the drawdown.

Specific discharge or specific flux. The rate of discharge per unit cross-sectional area measured at right angles to the direction of flow. It has the dimensions of velocity.

Specific run-off. The rate of discharge per unit drainage area. It has the dimensions of velocity and may be measured in l/sec. km^2.

Specific yield (saturation deficit). The ratio of (a) the volume of water that a rock, after being saturated, will yield by gravity to (b) its own volume. The specific yield is expressed as a decimal fraction or as a percentage. The specific yield is usually determined from the difference between the maximum moisture capacity and the specific retention.

Temporary perched ground water. Ground water in an aquifer underlain by unsaturated material and that contains water only part of the time.

Transpiration. A quantity of water absorbed by vegetation in the process of its growth. The rate of transpiration depends both on physical factors (evaporation of moisture through plant stomata) and biological factors.

Unconfined (free) ground water. Ground water in a water-table aquifer. The surface of the ground-water body is at atmospheric pressure.

Water balance. A ratio between the inflow and outflow of water within a certain area. The components of the natural water balance are precipitation, surface and ground-water inflow evaporation, transpiration, condensation and run-off (surface and subsurface).

Well yield. Volume of water that is withdrawn from a well per unit of time. The yield is expressed in m^3/sec, m^3/hour, gallons per minute or hour.

Zone of aeration. The upper zone of the earth's crust between the land surface and the water table. The pores, fissures and other interstices of the soils and rocks of the zone of aeration contain pellicular and capillary water. Gravitational water commonly percolates into and through this zone. Many voids are occupied by water vapour and other gases.

D Values of the function $f(\beta, z/L)$

(See Section 5.5.2, Figure 5.5.2b and Equation 5.5(3)

β_L, – z/L	0	0.1	0.2	0.3	0.4	0.5	0.6	0.7	0.8	0.9	1.0
0.001	0	0.1000	0.1999	0.3000	0.3999	0.4999	0.5999	0.6999	0.7999	0.9000	1.0000
0.01	0	0.0996	0.1992	0.2990	0.3988	0.4988	0.5988	0.6989	0.7992	0.8995	1.0000
0.10	0	0.0956	0.1921	0.2896	0.3880	0.4875	0.5680	0.6894	0.7919	0.8954	1.0000
0.25	0	0.0891	0.1805	0.2742	0.3703	0.4688	0.5898	0.6733	0.7795	0.8884	1.0000
0.50	0	0.0790	0.1621	0.2495	0.3413	0.4378	0.5393	0.6460	0.7581	0.8760	1.0000
0.75	0	0.0697	0.1449	0.2259	0.3132	0.4073	0.5088	0.6181	0.7360	0.8630	1.0000
1.0	0	0.0612	0.1289	0.2036	0.2862	0.3775	0.4785	0.5900	0.7132	0.8495	1.0000
2.0	0	0.0347	0.0770	0.1287	0.1918	0.2689	0.3631	0.4782	0.6187	0.7904	1.0000
3.0	0	0.0183	0.0431	0.0765	0.1216	0.1824	0.2646	0.3755	0.5252	0.7272	1.0000
4.0	0	0.0092	0.0229	0.0433	0.0738	0.1192	0.1870	0.2882	0.4391	0.6642	1.0000
5.0	0	0.0044	0.0117	0.0236	0.0433	0.7586	0.1295	0.2179	0.3636	0.6039	1.0000
6.0	0	0.0020	0.0058	0.0125	0.0249	0.4743	0.0885	0.1632	0.2995	0.5477	1.0000
7.0	0	0.0009	0.0028	0.0065	0.0141	0.2931	0.0600	0.1217	0.2459	0.4961	1.0000
8.0	0	0.0004	0.0013	0.0034	0.0079	0.1799	0.0404	0.0904	0.2016	0.4491	1.0000
9.0	0	0.0002	0.0006	0.0017	0.0044	0.1099	0.0272	0.0671	0.1652	0.4065	1.0000
10	0	0.0001	0.0003	0.0009	0.0024	0.0067	0.0183	0.0497	0.1353	0.3679	1.0000
15	0	0.0000	0.0000	0.0000	0.0001	0.0006	0.0025	0.0111	0.0498	0.2231	1.0000
20	0	0.0000	0.0000	0.0000	0.0000	0.0000	0.0003	0.0025	0.0183	0.1353	1.0000
30	0	0.0000	0.0000	0.0000	0.0000	0.0000	0.0000	0.0001	0.0025	0.0498	1.0000
100	0	0.0000	0.0000	0.0000	0.0000	0.0000	0.0000	0.0000	0.0000	0.0000	1.0000
–0.001	0	0.1000	0.2001	0.3001	0.4001	0.5001	0.6001	0.7001	0.8001	0.9000	1.0000
–0.01	0	0.1005	0.2008	0.3011	0.4012	0.5012	0.6012	0.7010	0.8008	0.9004	1.0000
–0.10	0	0.1046	0.2081	0.3106	0.4120	0.5125	0.6120	0.7104	0.8079	0.9044	1.0000
–0.25	0	0.1116	0.2205	0.3267	0.4302	0.5312	0.6297	0.7258	0.8195	0.9109	1.0000
–0.50	0	0.1240	0.2419	0.3540	0.4607	0.5622	0.6587	0.7505	0.8379	0.9210	1.0000
–0.75	0	0.1369	0.2640	0.3819	0.4912	0.5927	0.6868	0.7741	0.8551	0.9303	1.0000
-1.0	0	0.1505	0.2868	0.4100	0.5215	0.6225	0.7138	0.7964	0.8711	0.9388	1.0000
-2.0	0	0.2096	0.3813	0.5218	0.6369	0.7311	0.8082	0.8713	0.9230	0.9653	1.0000
-3.0	0	0.2728	0.4748	0.6245	0.7354	0.8176	0.8784	0.9235	0.9569	0.9817	1.0000
–4.0	0	0.3358	0.5609	0.7118	0.8130	0.8808	0.9262	0.9567	0.9771	0.9908	1.0000
-5.0	0	0.3961	0.6364	0.7821	0.8705	0.9241	0.9567	0.9764	0.9883	0.9956	1.0000
-6.0	0	0.4523	0.7005	0.8368	0.9115	0.9526	0.9751	0.9875	0.9942	0.9980	1.0000
-7.0	0	0.5039	0.7541	0.8783	0.9400	0.9707	0.9859	0.9935	0.9972	0.9991	1.0000
-8.0	0	0.5509	0.7984	0.9096	0.9596	0.9820	0.9921	0.9966	0.9987	0.9996	1.0000
-9.0	0	0.5935	0.8348	0.9329	0.9728	0.9890	0.9956	0.9983	0.9994	0.9998	1.0000
-10	0	0.6321	0.8647	0.9503	0.9817	0.9933	0.9976	0.9991	0.9997	0.9999	1.0000
-15	0	0.7769	0.9502	0.9889	0.9975	0.9994	0.9999	1.0000	1.0000	1.0000	1.0000
-20	0	0.8647	0.9817	0.9975	0.9997	1.0000	1.0000	1.0000	1.0000	1.0000	1.0000
-30	0	0.9502	0.9975	0.9999	1.0000	1.0000	1.0000	1.0000	1.0000	1.0000	1.0000
-100	0	1.0000	1.0000	1.0000	1.0000	1.0000	1.0000	1.0000	1.0000	1.0000	1.0000

E Values of the well function, $W(u)$, for values of u between 10^{-15} and 9.5

(See Section 6.1.1.4. and Equations 6.1 (17) and 6.1 (18))

N \ u	$N \times 10^{-15}$	$N \times 10^{-14}$	$N \times 10^{-13}$	$N \times 10^{-12}$	$N \times 10^{-11}$	$N \times 10^{-10}$	$N \times 10^{-9}$	$N \times 10^{-8}$	$N \times 10^{-7}$	$N \times 10^{-6}$	$N \times 10^{-5}$	$N \times 10^{-4}$	$N \times 10^{-3}$	$N \times 10^{-2}$	$N \times 10^{-1}$	N
1.0	33.9616	31.6590	29.3564	27.0538	24.7512	22.4486	20.1460	17.8435	15.5409	13.2383	10.9357	8.6332	6.3315	4.0379	1.8229	0.2194
1.5	33.5561	31.2535	28.9509	26.6483	24.3458	22.0432	19.7406	17.4380	15.1354	12.8328	10.5303	8.2278	5.9266	3.6374	1.4645	.1000
2.0	33.2684	30.9658	28.6632	26.3607	24.0581	21.7555	19.4529	17.1503	14.8477	12.5451	10.2426	7.9402	5.6394	3.3547	1.2227	.04890
2.5	33.0453	30.7427	28.4401	26.1375	23.8349	21.5323	19.2298	16.9272	14.6242	12.3220	10.0194	7.7172	5.4167	3.1365	1.0443	.02491
3.0	32.8629	30.5604	28.2578	25.9552	23.6526	21.3500	19.0474	16.7449	14.4423	12.1397	9.8371	7.5348	5.2349	2.9591	.9057	.01305
3.5	32.7088	30.4062	28.1036	25.8010	23.4985	21.1959	18.8933	16.5907	14.2881	11.9855	9.6830	7.3807	5.0813	2.8099	.7942	.006970
4.0	32.5753	30.2727	27.9701	25.6675	23.3649	21.0623	18.7598	16.4572	14.1546	11.8520	9.5495	7.2472	4.9482	2.6813	.7024	.003779
4.5	32.4575	30.1549	27.8523	25.5497	23.2471	20.9446	18.6420	16.3394	14.0368	11.7342	9.4317	7.1295	4.8310	2.5684	.6253	.002073
5.0	32.3521	30.0495	27.7470	25.4444	23.1418	20.8392	18.5366	16.2340	13.9314	11.6289	9.3263	7.0242	4.7261	2.4679	.5598	.001148
5.5	32.2568	29.9542	27.6516	25.3491	23.0465	20.7439	18.4413	16.1387	13.8361	11.5336	9.2310	6.9289	4.6313	2.3775	.5034	.0006409
6.0	32.1698	29.8672	27.5646	25.2620	22.9595	20.6569	18.3543	16.0517	13.7491	11.4465	9.1440	6.8420	4.5448	2.2953	.4544	.0003601
6.5	32.0898	29.7872	27.4846	25.1820	22.8794	20.5768	18.2742	15.9717	13.6691	11.3665	9.0640	6.7620	4.4652	2.2201	.4115	.0002034
7.0	32.0156	29.7131	27.4105	25.1079	22.8053	20.5027	18.2001	15.8976	13.5950	11.2924	8.9899	6.6879	4.3916	2.1508	.3738	.0001155
7.5	31.9467	29.6441	27.3415	25.0389	22.7363	20.4337	18.1311	15.8286	13.5260	11.2234	8.9209	6.6190	4.3231	2.0867	.3403	.00006583
8.0	31.8821	29.5795	27.2769	24.9744	22.6718	20.3692	18.0666	15.7640	13.4614	11.1589	8.8563	6.5545	4.2591	2.0269	.3106	.00003767
8.5	31.8215	29.5189	27.2163	24.9137	22.6112	20.3086	18.0060	15.7034	13.4008	11.0982	8.7957	6.4939	4.1990	1.9711	.2840	.00002162
9.0	31.7643	29.4618	27.1592	24.8566	22.5440	20.2514	17.9488	15.6462	13.3437	11.0411	8.7386	6.4368	4.1423	1.9187	.2602	.00001245
9.5	31.7103	29.4077	27.1051	24.8025	22.4999	20.1973	17.8948	15.5922	13.2896	10.9870	8.6845	6.3828	4.0887	1.8695	.2387	.00007185

F

The complementary error function, $i^2\text{erfc}\ \lambda$

(See Section 6.2.1 and Equation 6.2(3))

λ	$i^2\text{erfc}\ \lambda$	λ	$i^2\text{erfc}\ \lambda$	λ	$i^2\text{erfc}\ \lambda$
0.00	0.2500	0.29	0.1239	0.66	0.0436
0.01	0.2444	0.30	0.1207	0.68	0.0410
0.02	0.2438	0.31	0.1176	0.70	0.0382
0.03	0.2335	0.32	0.1145	0.72	0.0362
0.04	0.2282	0.33	0.1116	0.74	0.0340
0.05	0.2230	0.34	0.1087	0.76	0.0319
0.06	0.2179	0.35	0.1058	0.78	0.0299
0.07	0.2129	0.36	0.1030	0.80	0.0280
0.08	0.2030	0.37	0.0998	0.82	0.0262
0.09	0.2031	0.38	0.0976	0.84	0.0246
0.10	0.1937	0.39	0.0950	0.86	0.0230
0.11	0.1937	0.40	0.0925	0.88	0.0215
0.12	0.1892	0.41	0.0900	0.90	0.0201
0.13	0.1847	0.42	0.0875	0.92	0.0187
0.14	0.1803	0.43	0.0852	0.94	0.0175
0.15	0.1760	0.44	0.0828	0.96	0.0163
0.16	0.1718	0.45	0.0806	0.98	0.0152
0.17	0.1676	0.46	0.0783	1.0	0.0142
0.18	0.1635	0.47	0.0762	1.1	0.0099
0.19	0.1596	0.48	0.0740	1.2	0.0068
0.20	0.1557	0.49	0.0720	1.3	0.0046
0.21	0.1518	0.50	0.0700	1.4	0.0030
0.22	0.1481	0.52	0.0661	1.5	0.0020
0.23	0.1444	0.54	0.0623	1.6	0.0013
0.24	0.1408	0.56	0.0583	1.7	0.0008
0.25	0.1373	0.58	0.0555	1.8	0.0005
0.26	0.1338	0.60	0.0523	1.9	0.0003
0.27	0.1304	0.62	0.0492	2.0	0.0002
0.28	0.1271	0.64	0.0463		

G

Standard units of deviation, *t*, and computed values of $t/\sqrt{n}$

(See Section 7.4.5)

Number of observations, n	$\sqrt{n}$	Fiducial limits (per cent)							
		20		10		5		1	
		t	$t/\sqrt{n}$	t	$t/\sqrt{n}$	t	$t/\sqrt{n}$	t	$t/\sqrt{n}$
2	1.41	3.08	2.18	6.31	4.48	12.71	9.02	63.66	45.10
3	1.73	1.89	1.09	2.92	1.69	4.30	2.48	9.92	5.73
4	2.00	1.64	0.82	2.35	1.18	3.18	1.59	5.84	2.92
5	2.24	1.53	0.68	2.13	0.95	2.78	1.24	4.60	2.05
6	2.45	1.48	0.60	2.01	0.82	2.57	1.05	4.03	1.65
7	2.65	1.44	0.54	1.94	0.73	2.44	0.92	3.71	1.40
8	2.83	1.41	0.50	1.89	0.67	2.36	0.84	3.50	1.24
9	3.00	1.40	0.47	1.86	0.62	2.31	0.77	3.35	1.12
10	3.16	1.38	0.44	1.83	0.58	2.26	0.72	3.25	1.03
11	3.32	1.37	0.41	1.81	0.55	2.23	0.67	3.17	0.95
12	3.46	1.36	0.39	1.80	0.52	2.20	0.64	3.11	0.90
13	3.61	1.36	0.38	1.78	0.49	2.18	0.60	3.06	0.85
14	3.74	1.35	0.36	1.77	0.47	2.16	0.58	3.01	0.80
15	3.87	1.34	0.35	1.76	0.46	2.14	0.55	2.98	0.77
16	4.00	1.34	0.34	1.75	0.44	2.13	0.53	2.95	0.74
17	4.12	1.34	0.32	1.75	0.42	2.12	0.51	2.92	0.71
18	4.24	1.33	0.31	1.74	0.41	2.11	0.50	2.90	0.68
19	4.36	1.33	0.31	1.73	0.40	2.10	0.48	2.88	0.66
20	4.47	1.33	0.30	1.73	0.39	2.09	0.47	2.86	0.64
21	4.58	1.32	0.29	1.72	0.38	2.09	0.46	2.84	0.62
22	4.69	1.32	0.29	1.72	0.37	2.08	0.44	2.83	0.60
23	4.80	1.32	0.28	1.72	0.36	2.07	0.43	2.82	0.58
24	4.90	1.32	0.27	1.71	0.35	2.07	0.42	2.81	0.57
25	5.00	1.32	0.26	1.71	0.34	2.06	0.41	2.80	0.56
26	5.10	1.32	0.26	1.71	0.33	2.06	0.40	2.79	0.54
27	5.20	1.31	0.25	1.71	0.32	2.06	0.39	2.78	0.53
28	5.29	1.31	0.25	1.70	0.32	2.05	0.38	2.77	0.52
29	5.38	1.31	0.24	1.70	0.31	2.05	0.37	2.76	0.51
30	5.48	1.31	0.24	1.69	0.31	2.04	0.37	2.75	0.50
∞		1.28		1.64		1.96		2.58	