DEVELOPMENTS IN WATER SCIENCE

38

J. Balek

GROUNDWATER RESOURCES ASSESSMENT

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Groundwater Resources Assessment

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GROUNDWATER RESOURCES ASSESSMENT

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PREFACE

Many contemporary books on hydrogeology are based on advanced mathematical methods. Such an approach is needed for hydrology to develop further as a science. However, hydrogeological surveys are normally performed in rather complicated groundwater systems in which it may be difficult to define physical properties and boundary conditions. At the same time the environmental features of each region should be emphasized. In such cases deterministic and stochastic simulations have proved to be satisfactory. A water balance approach in both types of simulation provides very valuable information on the regional aspects of the solution.

C. C. Kisiel and L. Duckstein stated some years ago that "... the future effort in groundwater hydrogeology will transfer from theoretical approach to regional specialised problems and simulation of complex systems". An application of the system approach rules can help considerably in solving such regional problems. The principles of the system approach in water resources management were formulated by N. Buras and A. K. Biswas. Someone who is not familiar with water resources survey on a regional scale, may consider the systems approach rules as trivial. It is only when he attempts to follow some other approach that he will discover how many problems arise.

Once I was commissioned to set up a water balance model of a very extensive region. Because the application of standard methods would have taken many years, the investors accepted a proposal for the model approach, although this was considered as belonging to the sphere of extravagant hydrology. As usual, I tried to set up the system approach rules; however, I found that the project goals, subregional and even regional boundaries were far from being definite: in fact they were changed many times during the project. Instead of demanding a general strategy, a solution of minor details was requested, unimportant on a regional scale. One man linked with the project asked me to simulate a small catchment he happened to know, while another requested the same for a single borehole. The region was considered then a hydrological wonderland where soil profiles were fully saturated all the year round and the recharge rate varied only according to the altitude; of course I was requested to produce the model outputs in the same way. One man, particularly doubtful about any computer work, searched the scrapped printouts in an attempt to find a fatal error. He also requested confirmation of the model results by other than computerized methods: a job, which, when properly done, could well take many years.

Any attempt to explain the principles of system approach caused only indulgent smiles; those principles were considered a kind of project decoration which could be added to the final report in the very last stage.

First I was amazed, then bewildered, and finally amused. At the point I set up perhaps the first rules of the non-system approach: to generate unspecified results in regions with disputed boundaries, with no constraints on constraints. The solution to such a concept is not discussed in this book. However, if this book were to be dedicated to anyone, several people associated with that project would deserve it; they assisted in its genesis, though they were not aware of it.

Nevertheless, I am sincerely indebted to Mrs. K. Martonová and Mr. M. Bursík for their comments and suggestions, and to Mrs. K. Šedivá for the illustrations and graphic work.

J. Balek

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1 RÔLE OF HYDROLOGY IN GROUNDWATER RESOURCES MANAGEMENT

Groundwater is one of the most important natural resources. In semi-arid and arid regions it is often the only source of water, and in humid regions it plays an equally important rôle since surface water sources are frequently polluted. In some humid regions surface water resources are characterized by intermittency, scarcity, uneven distribution and inadequate development. Establishing a low-cost groundwater supply serves social and economical needs without endangering the environment.

In many parts of the world a survey of groundwater resources is spontaneous rather than systematic. Among the many wells and boreholes drilled only a few have been systematically observed for a prolonged period of time, and few records of groundwater level fluctuation are available for periods longer than ten years.

The present trend in groundwater utilisation tends to distinguish between renewable and non-renewable resources, and leads to the determination of maximum and minimum water levels in order to regulate the storage capacity. Also, the water balance approach in groundwater resources assessment must be examined from the point of view of the time delay between effective precipitation and groundwater outflow into the stream network.

The effect of man's activities should be borne in mind in order to prevent unfavourable economic conditions which could result in damage to aquifers. Reports from various parts of the world of a steady decline in groundwater levels are alarming and, what is worse, the damage inflicted in many cases may be irreversible. A decline of groundwater level due to excessive pumping causes permanent damage to the stability of groundwater storage, increases the cost of further pumping, and facilitates seawater intrusion into coastal aquifers.

The process of groundwater storage enrichment and depletion has to be viewed as an integral part of complex interactions between the atmosphere, vegetation, soil, aquifer and river networks.

Apart from the historically exogenous groundwater sources stored in aquifers during past geologic periods, the following main sources of groundwater recharge are recognized: Condensation of atmospheric humidity, rainfall, and infiltration of surface water from streams and lakes.

The first source is less significant than the recharge from rainfall, particularly during those periods when rainfall coincides with increased humidity and saturated soil profile; otherwise the recharge is greatly reduced by evapotranspiration. An infiltration of water from surface sources is sometimes promoted artificially in an attempt to improve existing ground water resources.

An important rôle is played by rock formations and vegetational cover. A bare rock containing fissures may significantly increase the recharge, whereas vegetation may reduce it directly or through the transpiration of stored groundwater. Very little attention has been devoted to the preferred pathways which, according to the opinion of many hydrogeologists, are a dominant feature of many water-bearing systems.

These phenomena are often neglected in the model approach towards the solving of groundwater problems. Many models have very little respect, if any, for a given hydrogeological environment and aquatic ecosystem. System approach and system analysis help the model-maker to identify and select a preferred model, among several alternatives, which is able to provide insight into the problem. The choice of a simulation routine is nearly always a compromise between mathematical accuracy and description of the real-world situation.

Groundwater exploitation is an important topic in modern hydrogeology. The priority problems are often discussed. According to the proposals of UNDP, the priority in utilizing groundwater resources should be given to projects for the supply of water for drinking and domestic purposes, second priority is given to water supply for cattle and herds, followed by supplies for irrigation purposes and industry.

In the developing world obviously water for agriculture and rural water supply is high on the priority list. Because the population in many developing countries is growing faster than the provision of food, an improved water supply to the fields is one of the most efficient means of increasing food production. A considerable improvement can be achieved through effective irrigation, which is recommended in arid lands where the settled agriculture depends on irrigation, semi-arid lands to ensure safe yields, areas with erratic rainfall for the production of high-value export crops, and areas suitable for crops with higher water requirements.

It is typical for most of these areas that local water is available in small supplies, such as pan valleys, and from groundwater resources. At first sight groundwater utilisation for such a purpose may appear very uneconomical, but in some regions, for instance in Africa, areas with precipitation between 0 – 200 mm have some of the world's greatest water-bearing aquifers. In the Nubian sandstones and Continental Intercalary Fountain are stored about $600 \cdot 10^3\,km^3$ of water. It has been estimated that by utilising $10\,m^3s^{-1}$ of Sahara's fossil waters, no more than one-thousandth of the resources would be consumed in twenty years. Natural or drilled artesian water outflows forming lakes and swamps, known as shebkas in Libya and shoots in Algeria, have been occasionally tapped manually by local inhabitants down to depths of almost 150 metres.

Extensive projects of groundwater utilisation for agriculture are reported from

Libya and Iran. In western and southern Australia, in the absence of better groundwater resources, cavities in granitic formations known as gnamma are explored. In New South Wales attention is paid to the groundwater in flat regions flooded during the periods of erratic rainfall. A further increase in groundwater consumption for agriculture is foreseen. Unfortunately, the sources are often limited and must be always carefully examined long before any exploitation begins.

In urban areas two major problems can be observed, namely how to ensure an adequate supply of water and how to dispose of the used water. It is a world-wide misconception to regard the disposal problem as less important than that of water supply. A moderate, steady increase in demand for water can usually be handled in a city where an equilibrium is maintained between the number of citizens, power supply, available sources of water and the size of the water distribution system. In many cities the municipal authorities are unable to keep control over the water supply because of uncontrolled migration of people from rural areas. Standpipes represent a temporary solution, but an additional system problem is the maintenance of old piping systems.

The growth of cities creates a serious effect upon the hydrological cycle. While in a natural catchment the groundwater resources are replenished at more or less regular intervals by atmospheric water, this is not so easy in urbanised areas where much vegetation has been removed to make room for construction sites, streets and engineering networks, and where paving of the surface has changed the infiltration rates. The water quality will also be affected to a considerable extent in many cases.

Water recycling in the vicinity of urbanised areas has become an economic solution. A better knowledge of the soil-groundwater-stream interactions is important for the design of such recycling systems.

Although groundwater contributes only marginally to the industrial water supply, industrial and mining activities have a tremendous effect on groundwater resources. Among the most important problems at present are: the eutrophication of natural waters and their microchemical pollution by synthetic organic substances, detergents, phenols, thermal pollutants, hydrocarbons and nitrates.

From the hydrological point of view, the feasibility of industrial development in an area should first be based on an inventory of water resources, including groundwater. The availability of water must also be examined in the light of technologies currently available. Here the hydrological approach becomes more complicated, because the optimization of water use has to be analysed so as not to interfere with the pre-established economic goals which take into account the physical, political and sociological constraints of the region.

The decision whether a raindrop will contribute to groundwater resources and perhaps to the formation of a baseflow is made in the atmosphere, in the vegetation, and in the soil profile, days, weeks, years and sometimes millenia before the baseflow is actually formed. Therefore in this book an attempt is made to analyse groundwater systems as part of a naturally balanced environment. Common problems of groundwater resources assessment and working methods have been minimized when information can be obtained from other sources, listed at the end of each chapter.

2 HYDROLOGICAL APPROACH IN GROUNDWATER RESOURCES ASSESSMENT

2.1 HYDROLOGICAL AND HYDROGEOLOGICAL REGIONS

While a region in general is defined as a territory of indefinite extent, in hydrology a region is defined as an area statistically homogeneous from a certain hydrological point of view (Dalrymple [1]). The flood régime, mean annual discharge and droughts are among the hydrological phenomena chosen for hydrological regionalisation. It does not matter which point of view we select; the objective of many hydrological methods is to obtain information which can be considered to be uniform within the boundaries of a certain region. In many cases, however, only the so called point data are available for regionalisation.

In a broader sense we talk about very extensive regions; for instance, tropical regions (Balek [2]) are limited by the parallels 23°27′ south and north of the Equator. Between them the sun is directly overhead twice a year and water circulation is governed by the direct and intensive influence of the sun's activity. These meteorological effects, however, are distorted by extra influences, and thus the boundaries of the tropics are sometimes taken to be the dividing lines between the easterly and westerly windsbelts. Nevertheless, inside such broad limits we can find a great variety of hydrological, hydrogeological and other régimes and thus further subregionalisation is often found feasible.

A hydrogeological region is considered to be an area with uniform hydrogeological properties. Formerly, the geological structure of the underlying aquifers was considered as decisive; today the groundwater level fluctuation, hydraulic properties of the soil and aquifer, and surface geometry are taken into account. Water quality and the possibility of groundwater exploitation are often considered also. For instance, as early as 1914 [3], the geological aspects were emphasised by Otockij who classified the hydrogeological regions of the European part of the USSR. The classification was extended by Vasilevskij [4], Lange [5] and Zajcev [6]. The basic principles were formulated by Ovchinnikov [7] who applied the hydrological approach, particularly the process of groundwater replenishment. In principle, short-term, seasonal and all-year replenishment was recognized. The rate of replenishment was also considered to be an important factor, in conjuction with the morphological and geological structure. Distinctions were considered between the groundwater régimes in the vicinity of divides, on slopes and on terraces.

In the regional hydrogeology of the United States there are frequent references to the groundwater areas or provinces. Initial attempts to classify these were linked with the names of Meinzer [8], [9] and Tolman [10]. The classification was related to the areal extent of the important superficial water-bearing and/or bedrock formation, and the quantity of water and its mineral content.

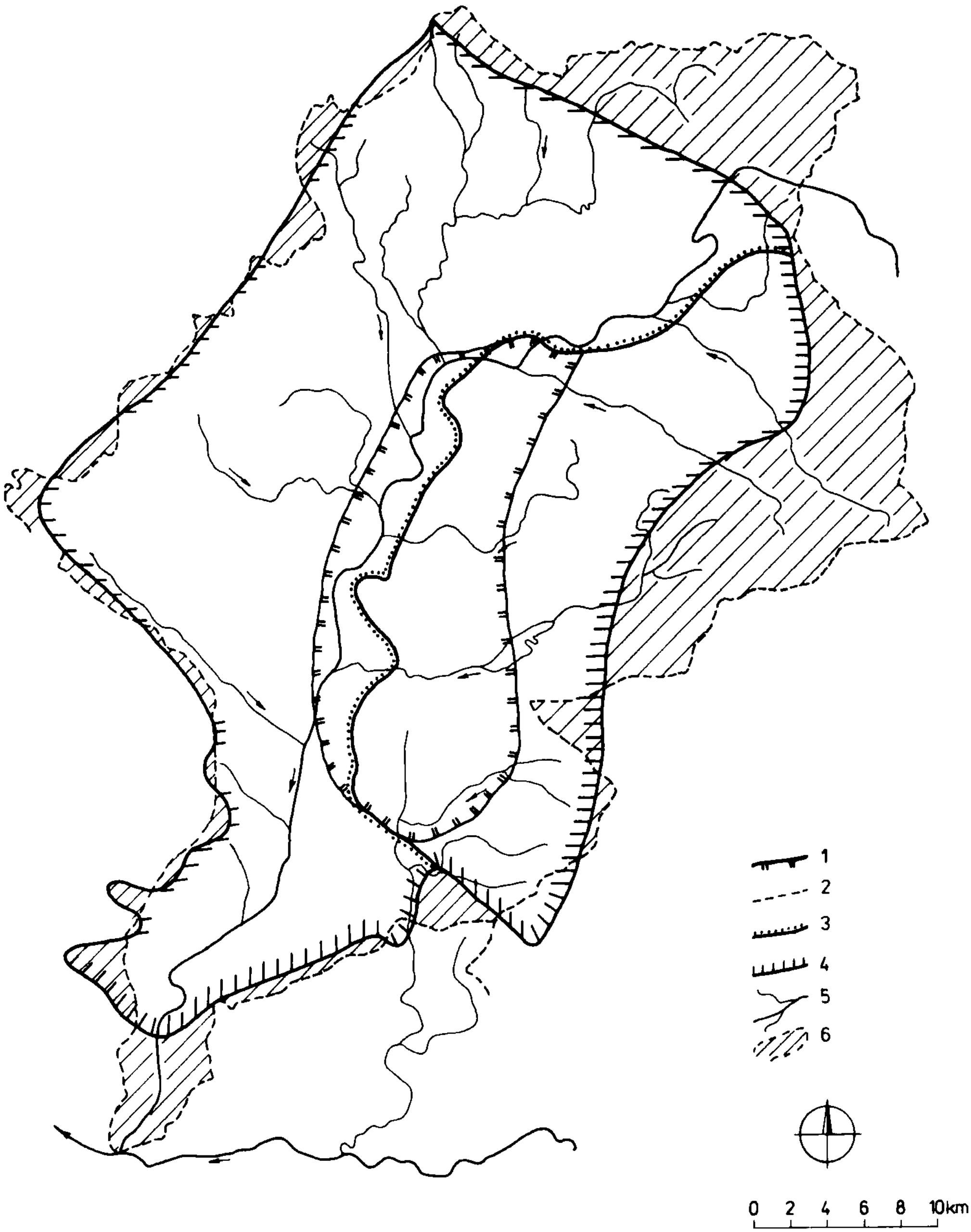
Sometimes the age of the water-bearing formations, their lithological development and tectonic structure, together with the water balance aspects are used to delineate hydrogeological regions (Franko et al. [11]).

A hydrologist tackling a hydrogeological problem will usually find it more feasible to assess the groundwater resources within an area defined by hydrological boundaries, where the water balance equation can be specified more clearly than in an area with many surface and subsurface inlets and outlets. Often, however, the boundaries of an area involved are defined in broader hydrogeological or administrative terms. This often assumed that when the hydrogeological features are uniform, the other properties, particularly the hydrological régime, will be also more or less uniform. However, on a hydrogeological map a region shown by a single colour may consist of a variety of subregions, each of them with different vegetation, soil and morphology. These subregions rarely have sharp boundaries; often there are very complex systems of interaction and superposition, and groundwater flow from one subregion to another may occur. Some interactions do not occur only between neighbouring regions. When the groundwater régime is simulated under such circumstances, the so-called boundary conditions have to be set up. An approximate determination of the boundary conditions may significantly reduce the mathematical accuracy of theoretically advanced solutions. Therefore it should always be made clear whether the boundary conditions of the region can be specified. If such a specification is not feasible, a simpler approach should be followed.

2.2 WATERSHED, CATCHMENT AND DRAINAGE BASIN

As stated by Ven Te Chow [12], the terms watershed and drainage basin are often considered to be synonymous. However, the term basin is reserved for more extensive drainage areas. In British terminology the term watershed is reserved for the divide separating one drainage basin from another, while a drainage area of small size is called a catchment. As can be seen in Fig. 2.1, the river basin of a large river may be composed of several hydrogeological regions or parts of them, and, vice versa, a uniform hydrogeological region can comprise several basins or parts of them.

Hydrological and hydrogeological boundaries frequently are not identical. Simple differences between hydrological and hydrogeological boundaries are well known; however, in some cases the situation can be complicated. In Fig. 2.2 are plotted the hydrological and hydrogeological boundaries of a catchment which is joined by the subsurface inflow of water piped along a fault from a distant source. Finding the boundaries of the hydrogeologically active region in such a case is



2.1 Simplified chart of the Cretaceous region in northern Czechoslovakia. Region with two water-bearing aquifers – *1*; hydrological divide – *2*; boundaries between Upper and Middle Turonian formations – *3*; hydrogeological divide – *4*; stream network – *5*; difference between hydrological and hydrogeological drainage area – *6*.

2.2 Hydrological (---) and hydrogeological (-·-·) divides of a catchment with long-distance communication (→).

difficult task, which must be solved by special identifying methods, such as the water balance, groundwater dating and water quality analysis. Even so, in many cases only approximate solutions can be achieved.

On a regional scale, groundwater basins are sometimes recognized. This term is reserved for an area with a closed circulation of groundwater. Such a basin can be formed by one or several aquifers connected vertically and/or horizontally. The connection may become active only after an extensive exploitation of one aquifer has been initiated.

The term "pan" is used less often for a closed groundwater circulation. The term "system" has become popular in simulation techniques. In general it is reserved for an aggregation of objects united by some form of regular interaction or interdependence. Not only aquifers are considered as objects; sometimes soil moisture zones and surface are considered equally important components of a system.

2.3 COMPONENTS OF THE WATER BALANCE EQUATION

In many hydrogeological studies we find the water balance equation in its simplest form, namely as a relationship between rainfall on one side of the equation and runoff and loss (sometimes incorrectly called evapotranspiration) on the other:

$$P = O + E_t \qquad 2.1$$

This equation contains very little information for groundwater régime analysis, because it gives no information about the ratio of the surface runoff and baseflow components. Also, while its validity is approximately correct for longtime periods, for shorter periods of time or single years the deficiencies caused by the soil moisture and groundwater storage increments and deficits are neglected. Therefore the following water balance equation becomes more informative:

$$P = R \pm O \pm \Delta G \pm \Delta S + I \pm C \pm M + E_t \qquad 2.2$$

Here P stands for precipitation, E_t for evapotranspiration, R for surface runoff, O for the groundwater outflow, G for the groundwater storage increment/deficit, S for the soil moisture increment/deficit, I for the amount of water intercepted, C for the communication with surrounding areas, M for water recharge or depletion caused by human activity; with all values in millimetres per period. Sometimes W component is included, being the water content of the accumulated snowpack.

In some situations even this extended equation may be found insufficient. This happens, for instance, when several aquifers of different behaviour contribute to the total outflow. Often, also, that part of the baseflow which flows through the shallow aquifers as the so-called subsurface runoff is omitted, although the baseflow from shallow aquifers forms an important part of many hydrographs. Furthermore, in complicated structures it should be recognised that the total runoff for a certain period of time is a mixture of water of different origins and times of residence. Therefore a general water balance equation should be written in a more complex form for a specified period:

$$P + \sum_1^j C_i = R + \sum_1^n O_r + \sum_1^n O'_r + \sum_1^m O_0 + \sum_1^m O'_0 \pm \sum_1^p M \pm$$
$$\pm \sum_1^{m+n} (\Delta G \pm \Delta S) + I + E_t \qquad 2.3$$

In addition to the previous components, here C_i represents the inflow of groundwater from another region, O_r is the outflow of the groundwater of recent origin through the stream network, O'_r is the outflow of the recent groundwater under the stream beds, O_0 and O'_0 are analogical components for the water of "old age"; and j, m, n, p indicate the subsystems/aquifers which contribute to the water balance, including shallow aquifers.

Such an equation can be written for an arbitrary time interval; however, the shorter the interval, the more difficult it is to obtain the data, because in principle a higher accuracy of measuring each component is required for shorter time intervals.

The water balance approach can vary according to the size of the balanced area. Basically, we can differentiate between:

a) a unit of rather small size characterised by a uniform groundwater régime, soil structure, vegetational pattern and orography,

b) a catchment defined by hydrological or slightly deviated hydrogeological boundaries,

c) a subregional or regional unit with boundaries differing from the hydrological ones, with non-uniform and complicated hydrological, hydrogeological, ecological and orographic patterns.

Small-size units are frequently featured in special research projects, for example as part of agricultural pilot schemes and instrument testing plots. The observations are concentrated within a small area, and can be easily maintained. However, the results have limited significance on a regional scale.

Also, the features of a single catchment located within the boundaries of a studied region are often not widely typical, particularly when the regional features are non-uniform. Therefore, some hydrologists conclude that the so-called "representative catchments" established in a region do not represent anything except themselves. Nevertheless, useful data have been obtained from representative catchments, for instance, for a comparison of the behaviour of various types of groundwater régimes.

The water balance assessment of a large region can be given by a single water balance equation in the same way as for a small unit or for a single catchment. However, one might ask what sort of information such an equation provides, particularly from the point of view of the behaviour of groundwater resources. For this purpose it can be more useful to set up several water balance equations for each subregion that can be considered as a single hydrological unit.

2.4 PRECIPITATION

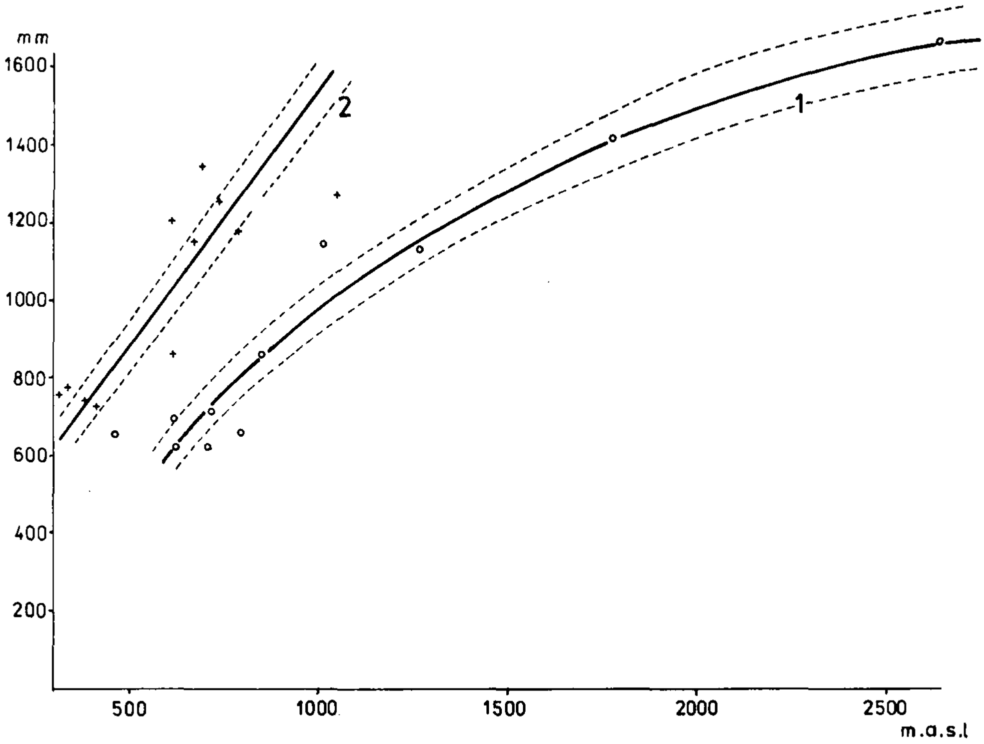
In many parts of the world not only rainfall but also snowfall and even mist are significant components of the water balance. Unlike rainfall, snow water is held for some time on the soil surface and vegetation and thus participates in the hydrological cycle with a certain delay. Snow accumulated during the winter period contributes more efficiently, after melting, to the formation of the conditions favourable for soil saturation and groundwater recharge. In temperate regions more complicated conditions of several snow accumulation/thawing periods occur, sometimes on frozen, sometimes on unfrozen soil.

Besides rain and snow, mist may be important. From a hydrometeorological viewpoint there is no basic difference between mist, fog and cloud. All are composed of small water droplets. If above ground, the name cloud is used, and when resting on the ground the terms mist or fog are used. In mist the range of visibility is between 1000 and 3000 metres, while thicker layers of droplets are called fog. These phenomena contribute to so-called horizontal precipitation, particularly under the effect of strong winds.

Using the water balance approach it has been calculated that, particularly in mountainous areas, up to 50% of the recorded annual rainfall should be added to represent the horizontal precipitation which is normally unrecorded. Fojt and Krečmer [13] found at the altitude of 900 m.a.s.l. in Northern Czechoslovakia that recorded annual rainfall should be increased at least by 20% to account for horizontal precipitation. Meteorological conditions favourable for the formation of horizontal precipitation can be traced by potential evaporation calculations at hourly intervals. Usually during those intervals for which negative values of the potential evaporation are obtained, the conditions have been favourable for the formation of horizontal precipitation.

Similarly, the effect of dew should be taken into account. A WMO guide [14] says that under favourable conditions there is a definite limit of 1.1 mm $hour^{-1}$ for the average rate of dew over an area. However, such a relatively small amount accumulated for a prolonged period can contribute significantly to the total precipitation budget. In the same publication, more information on precipitation measuring techniques and network design can be found.

In some regions we may conclude that the network of precipitation gauges is inadequate. Then, in the absence of direct measurements, the density of observation

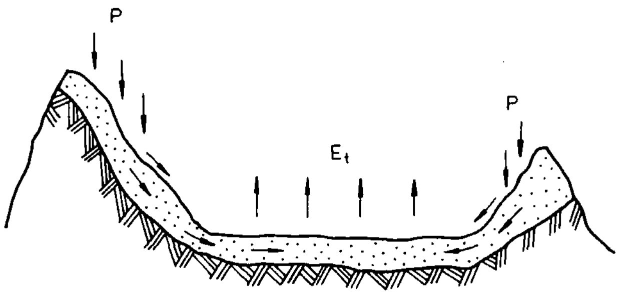


2.3 Altitude-rainfall relationship and the 95% credibility limits. Tatra Mountains – *1* and hilly areas in northern Czechoslovakia – *2*.

is increased by using the relationship between mean annual rainfall and the altitude (Fig. 2.3). It should be noted that such a relationship has only a regional validity and sometimes, in mountainous areas, we may find very different values for the windward and leeward sides of the mountains. Sometimes even a reverse effect has been observed, namely lower precipitation on the hilltops than in the valleys.

It is generally assumed that there is a close relationship between the groundwater level fluctuation and the precipitation régime. This assumption is valid when the soil profile is fully saturated. When the soil moisture content is well below the field capacity, a delay in the rise of groundwater level after rainfall can be expected.

Mean precipitation for a specified period of time over a basin or hydrogeological region is usually determined by the Thiessen polygon method, or by the isohyetal method. The Thiessen method assumes that the amount of rainfall at any station can be represented halfway to the surrounding stations. The isohyetal method consists of drawing lines of equal amount, using observed rainfall at stations and any additional factors such as topography to adjust or interpolate between observed stations. When employing this approach in model-making, it should not be forgotten that the groundwater level in an elementary area (defined in the model by a nodal system or by finite elements) actually may be a product of the precipitation which has occurred far away and a long time ago in the intake area. An example is the locally exogeneous groundwater source in a semi-arid pan (Fig. 2.4) where the groundwater storage is a product of more intensive rainfall on the slopes of the mountains,



2.4 Locally exogenous groundwater resources in a semi-arid pan. P – precipitation, E_t – evapotranspiration.

while the local rain, if any, falling on the pan is mostly or completely consumed by evapotranspiration. The groundwater replenishment can occur also in small areas of the region through preferred pathways; then the mean precipitation calculated for the whole region may be entirely different from the precipitation falling on those areas. Therefore the precipitation, in the water balance approach, should be considered rather as a contribution to the groundwater storage than a direct source of the baseflow.

2.5 RUNOFF PROCESS

In a general water balance approach, the runoff is represented by a single value corresponding to the mean discharge for the balanced interval. For groundwater resources assessment, separated baseflow and surface runoff have to be incorporated, together with the subsurface runoff component. The latter is of particular importance because it takes place over rather extensive areas during periods of potential recharge, when lateral flow prevents the water percolating to deep aquifers.

The first item to be specified in the water balance is the surface runoff. Traditionally, the surface runoff comprises that part of the runoff which flows over the surface to the stream. Also, the term overland flow was introduced by Horton [15]. In this simplified approach it was supposed to be a more or less uniform sheet of water flowing over the soil or, in other words, a process relevant to a uniform infiltration rate. Field experiments proved that the infiltration is far from being uniform and the surface runoff is subject mainly to areas of low infiltration (Dunin [16]). Thus "classical" overland flow can be observed in flat regions such as steppes and semi-arid to arid areas, while in temperate climates the surface runoff originates on paved roads, in urban areas, on bare soil, etc. Very little contribution can be expected from meadows, grassland and thick forest. Tab. 2.1 contains a comparison of the annual surface runoff from bare soil and grass.

Tab. 2.1 Surface runoff on grass and bare soil.
Experimental base Nedamov, Czechoslovakia.

Year	Grass mm	Bare soil mm
1982	37.53	58.27
1983	2.70	9.36
1984	1.96	26.54
Single storm:		
7. 8. 1982	0.20	6.07
One day snow melting and rain:		
1. 2. 1982	13.95	21.12

Rubin [17] analysed three modes of rain infiltration, according to the rainfall intensity and capillary conductivity. The so-called non-ponding, pre-ponding and ponded mode was recognised, and only the latter was assumed to be suitable for the formation of surface runoff.

In modern approaches the surface runoff theory is based on the partial area concept (Griend and Engmann [18]). The authors expect temporal and spatial variability of the contributing areas. Remote sensing methods facilitate the identification of those areas.

In shallow aquifers where the groundwater storage capacity can be exceeded and the lateral inflow into the aquifer continues, the surface runoff is formed from the bottom. Thus the formation of periodic intermittent swamps is explained (Balek, Perry [19]).

The groundwater runoff, also called groundwater outflow or baseflow, is produced by the water flowing from the aquifer into the stream network. In areas with perennial streams it is also called fair-weather runoff, because the hydrograph during prolonged periods without rainfall is formed by the groundwater sources. This becomes invalid soon after the rainy period, when the recession curve of the falling limb of the flood wave still contributes to the total discharge. Additionally, a prolonged effect of what is called the subsurface runoff (known also as subsurface flow, interflow, subsurface storm flow or slow seepage) is observed. According to our present knowledge, this part of the hydrograph represents the precipitation infiltrating into the surface soil layers and moving laterally through the upper horizons toward the streams in the form of shallow groundwater. In the vicinity of streams and in shallow pans the subsurface flow contributes to the river stream flow through shallow groundwater aquifers. Here it becomes temporarily deleted unless an active part of the stored groundwater begins to take an active role in contributing to the stream flow.

Sometimes a distinction is made between the prompt subsurface runoff and delayed subsurface runoff. Actually, within the basin boundaries a variety of prompt and delayed subsurface runoff may occur. This can be explained by the non-uniform entry properties of the surface and by varying travel times. Also the morphology, soil, vegetation, topography and other features play an important role (Betson [20]).

Baseflow, too, can be developed as a combination of the contribution from aquifers of various properties. They may contribute also to the non-uniform development of the intermediate groundwater in flow in different sections of the same stream.

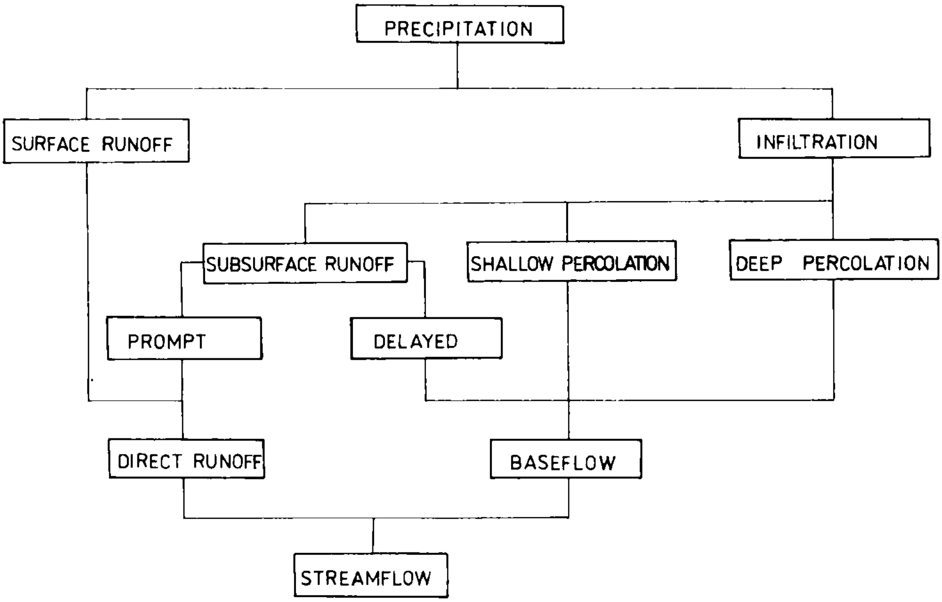
Conditions favourable for the formation of groundwater resources exist, for instance, in limestone and lava formations where the bedrock contains open fractures or lava tubes and soil cover is poorly developed. Here a very small fraction of precipitation contributes to the surface runoff, and the intermediate flow between two points of the longitudinal profile may vary from place to place. In some river sections the baseflow may even temporarily disappear, only to reappear again when the structure of the river bed permits. A similar effect can also be observed in regions where precipitation enters an aquifer which becomes confined, and water is transported far away to some other drainage area (Brown [21]).

The contribution of the groundwater flow to the stream can also be highly irregular when systems of faults and fissures cross the river bed.

Sometimes a high baseflow is observed due to reduced evapotranspiration in higher altitudes and on steep slopes. There the baseflow occurs as a fast response to precipitation. Usually this runoff pattern is unstable, and eventually diminishes during prolonged periods with low precipitation. Kněžek and Krásný [22] described this pattern in the Krkonoše Mountains on the Czechoslovak/Polish borders, where in the granitic formations of low permeability an extremely high baseflow yield has been observed. Troxel [23] found in Southern Californian formations that the steep, less permeable terraines yield a higher runoff than permeable areas for a given amount of precipitation.

A favourable effect from a hydrogeological point of view is found in areas with large fractures and highly permeable soils, where the water can quickly percolate into a deep aquifer before it can be consumed by evapotranspiration. On the other hand, in flat areas and shallow depressions the conditions are more conductive for reverse water transport to the atmosphere, and increased evapotranspiration is observed at the expense of groundwater recharge.

A simplified scheme of the runoff components is given in Fig. 2.5.



2.5 A simplified scheme of runoff components.

A detailed analysis of the intermediate baseflow development along the river can significantly contribute to the knowledge of the analysed groundwater system. Not only the discharge but also the yield is traced. The term "yield" in water balance studies is defined as

$$q \text{ litres s}^{-1}\text{ km}^{-2} = \frac{1000\,Q}{A}$$

where Q represents the discharge, $m^3 s^{-1}$, and A the drainage area, km^2. Obviously the runoff volume O

$$O \text{ mm year}^{-1} = q\,\frac{N}{1000^2}$$

where N is the number of seconds in a year.

The yield can be calculated for the total discharge or separately for each component, namely surface runoff, subsurface flow and baseflow. The separation of runoff components is closely related to the determination of groundwater resources and will be discussed in Chapter 3.

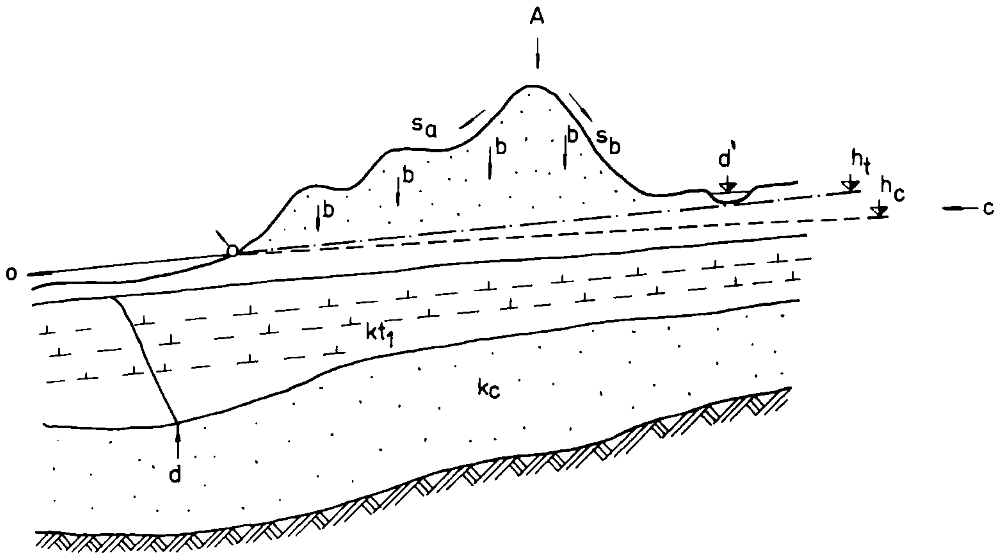
In groundwater hydrology the term "specific yield" also is common. It is the yield expressed as a percentage by volume of the drained formation. It is also defined as the volume of water that drains from an aquifer under the force of gravity, and, its value is obtained usually by analysing the pumping tests.

The term "delayed yield" may appear at first rather confusing in the hydrological context. Actually the term is relevant to the hydraulic analysis, and is found when drawdown plotted against time on the logarithmic scale forms an inflected curve consisting of a steep initial segment, followed by a flat segment at intermediate times, and a somewhat steeper segment at subsequent times (Neuman [24]). The intermediate segment suggests that some water is released from aquifer storage instantaneously as an additional source of water. Theories developed to explain this phenomenon have been related to the compressibility of even the unconfined aquifer, the release of water from the unsaturated zone, the air entry effect, etc. It does not appear to depend upon the time of residence in the aquifer.

2.6 COMMUNICATION COMPONENT

Under normal circumstances in an area defined by hydrological boundaries, the groundwater is recharged by infiltration. In complicated structures, concentrated inflow may exist from sources located outside the hydrological boundaries. This component is difficult to trace. Sometimes geophysical prospecting or well-logging methods can be very helpful. Chemical analysis of water samples, radioactive dating, and temperature measurements in the longitudinal profile have been applied with variable success. A water balance approach based on the analysis of the difference between observed values of water loss obtained from the water balance equations, and the water loss in the water balance calculated for surrounding catchments, will be discussed in Chapter 3.

In Fig. 2.6 is an example of the complexity of the communication as observed in Turonian formations. In a small tributary of the Labe river north of Prague, Czechoslovakia the baseflow is formed as a combination of: direct infiltration within the catchment's hydrological boundaries; direct infiltration outside those boundaries; and a long distance communication from the Turonian outcrops which quite probably do not have any common borders with the drainage area. In the lower reaches of the river, the catchment also communicates vertically with Cenomanian aquifer through fissures in the aquiclude. A combination of water balance and groundwater dating methods have been employed to assess each contribution to the formation of the baseflow.



2.6 An example of communications in Mesozoic formations. Cenomanian head – h_c, Turonian head – h_t, Turonian formations – kt_1, Cenomanian formations – k_c, vertical communication – d, catchment outflow – o, catchment divide – A, direct surface runoff contribution – s_a, indirect surface runoff contribution – s_b, fast recharge – b long distance communication – c, stream network outside – d'.

Long-distance communication is common in karst regions, though the necessary conditions may exist in any aquifer with a well-developed system of fissures and fractures. If the contribution to the balanced region from another distant region is significantly high, the communication component can be easily traced in the water balance approach. More often, however, the contribution from another aquifer is relatively small and its component in the water balance equation is rather approximate.

2.7 MAN'S IMPACT COMPONENT

Groundwater resources are subject to human impact in a variety of ways. Because surface water supplies are influenced by their seasonality, the exploitation of groundwater resources is increasing everywhere. In some places, in an attempt to prevent the groundwater level from declining further, or to create an adequate supply of water, artificial recharge is practised. Both pumping and artifical recharge have many secondary effects and therefore cannot be omitted from the water balance calculations. If, for instance, surface water is utilised for groundwater recharge, the ratio of the runoff components may change. If groundwater is pumped intensively from the vicinity of streams, a groundwater inducement effect is developed. Irrigation practised in areas with low precipitation and high potential evapotranspiration will change the vegetational pattern, and any resulting increase in evapotranspiration will cause a decline of the groundwater level (Fig. 2.7). And in fractured/fissured rocks, falling groundwater level caused by prolonged pumping may influence the régime of neighbouring aquifers to such an extent that the groundwater is tapped from them because of the changing gradient of the groundwater level.

In principle, three activities of man can be recognised (Colebrander [25]):

1. primary,
2. secondary,
3. inadvertent.

Primary activities are considered to include operational and engineering works undertaken with the aim of changing the hydrological régime or parts of it. With regard to the groundwater régime, these activities include artificial recharge of the aquifers, groundwater exploitation, drainage works, and river regulation. Usually, their consequences can be predicted quantitatively.

Secondary effects result from constructions and activities carried out for purposes other than groundwater recharge. Their secondary influence on the groundwater régime has to be taken into account. For instance, a dam built as part of water power scheme will cause a rise in groundwater level in the vicinity of the reservoir; an irrigation scheme may affect the groundwater recharge pattern by changing the vegetational cover. Waterlogging may be caused as a secondary effect of inefficient irrigation systems. The lining with concrete of water transport canals may cause the groundwater level to rise, etc.

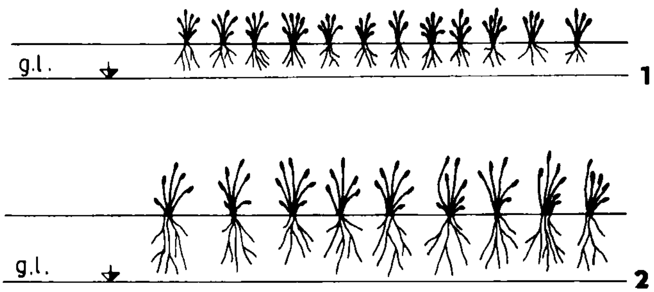
Inadvertent effects are those which at first sight may have nothing in common with the water régime, although they may affect at least parts of it more or less indirectly. Large-scale changes in land use may completely change the groundwater pattern. Forest fires are one example, and many human activities related to agriculture, forestry, urbanisation and industrialisation fall within this group.

Assuming groundwater changes to be a stochastic process, the following classification is applicable:

1. sudden changes,
2. gradual changes,
3. irregular changes.

Sudden changes are defined as those which occur immediately, such as those which follow the completion and putting into operation of water works. Rapid destruction of the vegetation by air pollutants or by fire also may initiate sudden changes in the groundwater régime.

Gradual changes are produced by long-lasting activities on a small scale. Protracted pumping of water in small quantities produces a slow but continuous downward movement of the groundwater level. A similar effect may be exerted by increasing urbanisation. Mining practices also may contribute to gradual changes, even though



2.7 Groundwater level before – *1* and after – *2* irrigation.

sudden catastrophes in the mines such as water bursts can cause irregular but sometimes catastrophic changes. Earthquakes, landslides and other natural catastrophes also may produce groundwater régime changes of an irregular character.

Irregular changes are the most difficult changes to include in water balance methods, whereas sudden and gradual changes can be more easily considered, even if with variable accuracy.

2.8 EVAPOTRANSPIRATION

In the water balance approach to solving various regional groundwater problems, evapotranspiration (in simplified water balance equations sometimes also called "water loss") is the component most difficult to measure. On the other hand this component greatly affects water content in the unsaturated zone and thus also the groundwater recharge. Actually, the so-called unsaturated zone becomes saturated when the evapotranspiration becomes negligible as compared with the precipitation.

Formerly, evapotranspiration was defined as a process by which water is evaporated from wet surface and transpired by plants (Ven Te Chow [12]). In such a definition, water evaporated from the intercepted storage on the leaf surface is considered as a part of the evapotranspiration. More generally, however, the evaporation from that part of the soil surface which is not covered by vegetation is considered also as part of the evapotranspiration. This may happen even if at first sight the soil surface is not actually wet. Evapotranspiration viewed as such a complex process is called "consumptive use" by some agriculturalists.

Transpiration is usually defined as the water exchange through plants between their root systems and the atmosphere. It is always related to the vegetation, and often only that water which is spent on the building of plant tissues is considered as transpiration.

Evaporation is a process by which water is changed from the liquid or solid state into the gaseous state. In a practical approach it is always related to the loss of water column from a free surface over a fixed time interval. Either direct observation, or a calculation based on the factors involved in the transfer of thermal energy in the evapotranspiration process, are employed.

Often the terms potential evaporation and potential evapotranspiration are introduced into the calculations. Both terms indicate the maximum amount of water which can be released to the atmosphere, providing the most favourable conditions persist throughout the time interval involved. In transpiration such conditions may exist, for a given meteorological situation, when the soil profile remains saturated at field capacity all the time.

Before the process of evapotranspiration is discussed from the point of view of groundwater regimes, the rôle of soil should be described.

2.8.1 Saturated and unsaturated soil zones

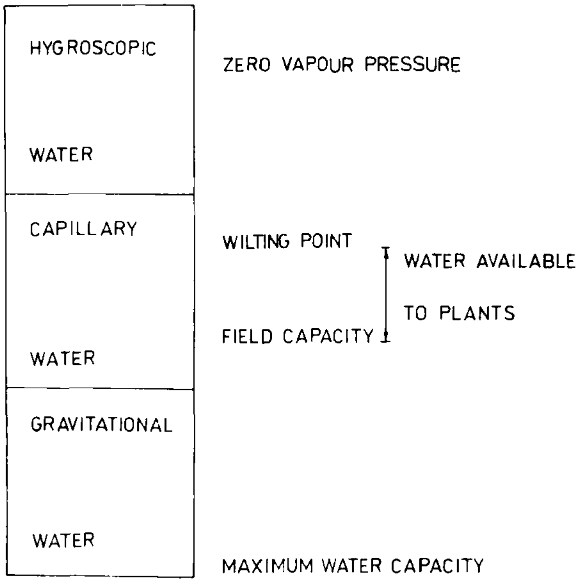
Soil is, in most cases, a significant medium of interaction between the atmosphere and vegetation on one hand, and groundwater on the other. Traditionally, from the hydrological point of view, two zones have been recognised: the zone of aeration (or the unsaturated zone) is considered to be that part of the soil profile which is, at least temporarily, partially filled with water and partially with air. Because the water is also stored in the capillary pores of the root systems of plants, the zone is sometimes called the surface-biological zone. This zone almost always overlies the zone of saturation and extends up to the soil surface.

The saturated zone (or the zone of saturation) can extend to the soil surface under special circumstances; alternatively it can be bounded by overlying strata. It extends down to the underlying impermeable strata. From a practical point of view, the unsaturated zone must become temporarily saturated in order for groundwater recharge to occur.

In the absence of impermeable strata the phreatic surface is defined as the upper surface of the zone of saturation. Thus, obviously, the saturated zone is also limited by the water level surface under the prevailing atmospheric pressure. Some soil physicists prefer to consider the saturated zone as containing also the zone of capillary rise, which extends above the water table owing to capillary forces. Water is held in this zone at less than atmospheric pressure.

Because the formation of the zone of aeration is so closely linked with vegetation, in some publications this zone is defined as that part of the soil which extends down through the root zone. However, in some aquatic ecosystems roots are so adapted that they can tap water from the groundwater resources through the zone of capillary rise. Because this zone is almost always saturated and recharged from the groundwater sources, it fluctuates depending on the groundwater level; occasionally it may actually contact the roots and it is therefore considered to be another medium between groundwater and the unsaturated zone.

Three categories of soil water can be recognised in the soil zones: Adsorbed or hygroscopic water forms a thin film of moisture on soil particle surfaces. Because of great adhesive force this water is not available to plants. Capillary water exists



2.8 Basic soil water categories.

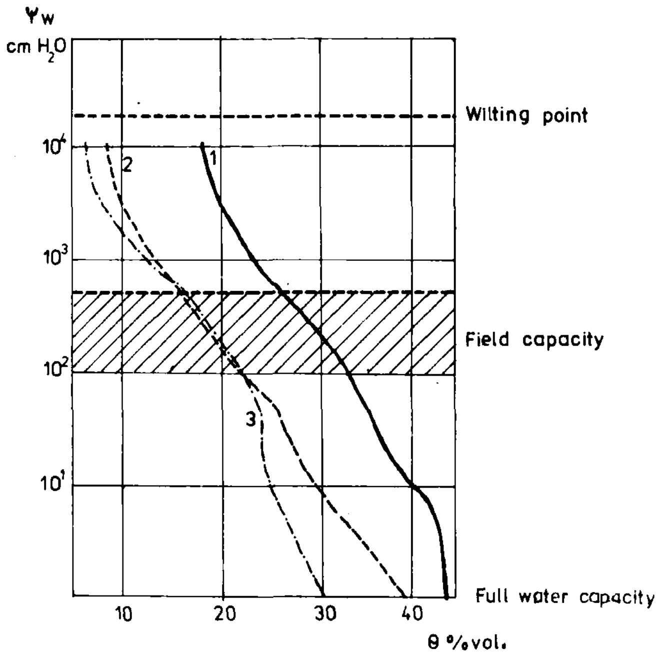
as a continuous film around soil particles. This water is moved by capillary forces and is available to plants and for soil evaporation. Gravitational water is capable of downward movement by gravitational force, and is actually the only water which can recharge the aquifer. The downward movement of water occurs only when the capillary pores have been filled. Thus capillary water is equally important in the process of groundwater recharge (Fig. 2.8).

The complex impact of forces on the soil water is called the capillary water potential. The term pF was introduced as to represent of the relationship between the suction pressure and soil moisture. Symbol p indicates the logarithmic relationship, the symbol F stands for the energy of enthalpy (given as the difference between free enthalpy of water at given soil moisture and free enthalpy of water at the water level surface of the soil-water system). In Fig. 2.9 there is an example of the pF curves for sandy loam. Here the suction pressure Ψ_w is in cms of water.

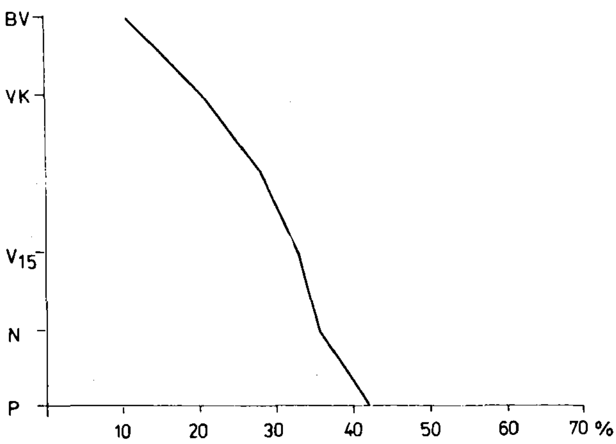
A comparison of pF curves for various types of soil can be utilised in the simulation when groundwater recharge areas are regionalised and classified.

In the absence of pF curves so-called hydrolimits can serve this purpose. In contrast with pF curves, simple laboratory equipment is sufficient to determine hydrolimits. An example of hydrolimits is given in Fig. 2.10 for the same soil as in Fig. 2.9. Here P stands for total porosity, N for absorption capacity, V_{15} is the volume of noncapillary pores, and VK is the retention capacity (also called absolute water capacity) defined as:

$$VK = \sqrt{(I + 18)\,20} \qquad \%\text{ of volume} \qquad 2.4$$



2.9 A variety of pF curves for different depths of sandy loam. Depth 10 cm – *1*, 50 cm – *2*, 110 cm – *3*.



2.10 Hydrolimits for sandy loam, depth 90 – 100 cm.

where I is the percentage of claylike particles of size below 0.1 mm. BV is the wilting point, defined as:

$$BV = 0.3I + 4 \qquad \% \text{ of volume} \tag{2.5}$$

If grain-size analysis is not available, I can be determined approximately (Tab. 2.2). Field capacity is defined as the amount of water held in the soil after the excess gravitational water has drained away and the rate of downward movement of water has materially decreased. Wilting point is the moisture content of soil at which it cannot supply water at a sufficient rate to maintain turgidity, and the plant wilts.

It is much more convenient to consider the additional effect of roots, (the atmosphere-soil-plant system), so important for the process of groundwater recharge.

Tab. 2.2 Percentage of clay-size particles in soils

Soil	Percentage of clay-size particles, %
Sand	0–10
Sandy loam	10–20
Sandy clay loam	20–30
Loam	30–45
Clay loam	45–60
Clay	60–90

The movement of soil water through a plant is described as a process which depends on the chemical potential or, in other words, on the water deficiency within a plant (Balek and Pavlík [26]). Molecules of water tend to move from sites at which they have a higher energy content to points where the mean energy content is lower. According to Fick's law the diffusion rate of a substance depends on: the gradient of water potential; the membrane cross-section; the temperature; and the concentration.

The water potential corresponds to the relative chemical potential so that:

$$\Psi = \mu - \mu_0 \qquad \mathrm{J\,m^{-3}} \tag{2.6}$$

where μ is the chemical potential of the water at any point in the system and μ_0, is the chemical potential of pure free water. The water potential consists of three independent components: a pressure component, a matrix component and an osmotic component, according to the following relation:

$$\Psi = \Psi_p + \Psi_m + \Psi_\pi \tag{2.7}$$

The first component, called the pressure potential, represents the actual pressure on the cell and is usually positive in the plants even though it may be negative in the cells. The matrix component represents that part of specific free energy which is related to the water in the colloidal structures of the cells and the capillary structure of the soil, and is always negative. In general, osmosis can be considered to be a thermodynamic process. The solvent, which has a higher vapour tension than the solution, tends to move towards the solution by osmosis, or isothermal distillation, or diffusion. The water potential gradient is also influenced by the temperature difference between two points in the plant.

The chemical potential can be expressed also by the equation

$$\Psi_w = RT \ln\left(\frac{p}{p_0}\right) \qquad \text{J mole}^{-1} \qquad 2.8$$

R is the gas constant, in J mole^{-1} K^{-1}; T is value of the absolute temperature, K; p is the water pressure and p_0 is the maximum water vapour pressure at a given temperature. This means that the water potential decreases with a decrease in the vapour pressure, the latter corresponding to the relative air humidity. As early as 1931 , Walter [28] introduced the term hydrature as the relative water vapour pressure, or the maximum water vapour tension at the given temperature.

While the problem of water movement in plants has been studied theoretically from the beginning of the century (Sutcliffe [27]), so far no uniform opinion has been reached among biophysicists on the mechanics of the whole process, and neither has the rôle of fluctuating groundwater level been taken into account. Measurements of the water potential under laboratory conditions are well described (Slavík et al. [29]), but, when attempts are made to transfer laboratory results to the field, numerous problems arise.

The theoretical conclusions reached by plant physiologists can serve at least as a guide to the selection of the factors which possibly influence sap movement in plants, and thus also influence the transpirational process.

As far as the soil water flux is concerned, a distinction must be made between saturated and nonsaturated soils. In saturated soils, Darcy's law is applied, so that the soil water flux, Q, is given

$$Q = \frac{k(h_i - h_b)\,A}{L} \qquad \text{m}^3\ \text{day}^{-1} \qquad 2.9$$

where L is the length of the soil column, m; A is the cross-section of the column through which the volume of water moves per day, in m^3; h_i is the head at the top of the column, m; h_b is the head at the bottom of the column, m; and k is the hydraulic conductivity, related to the soil permeability k' so that

$$k = \frac{k'\varrho g}{\eta} \qquad 2.10$$

where ϱ is the density of water, g the acceleration due to gravity and η is the viscosity. The velocity of water is

$$v = \frac{Q}{A} \qquad 2.11$$

and the hydraulic gradient

$$I = \frac{h_i - h_b}{L} \qquad 2.12$$

Therefore Darcy's law can be written in the simpler form

$$v = kI$$

It should be noted that because I is the hydraulic gradient and not the pressure gradient, for zero pressure we obtain

$$Q = \frac{kLA}{L} \qquad 2.13$$

and

$$\frac{Q}{A} = v = k \qquad 2.14$$

In other words, the zero pressure gradient does not correspond to zero flow.

According to Poiseuille's law, velocity in laminar flow is proportional to the first power of the hydraulic gradient, and it is assumed that Darcy's law is valid for laminar flow in porous media.

For flow in non-porous media, the Reynold's number serves as a criterion to distinguish between laminar and turbulent flow:

$$N_R = \frac{\varrho v D}{\eta} \qquad 2.15$$

where ϱ is the fluid density, v is the velocity, D the diameter of the medium and η the viscosity of the fluid. Departures from a linear relationship appear when

$$1 < N_R < 10$$

Departures from Darcy's law are found in rock formations, in unconsolidated aquifers which have steep hydraulic gradients, and in karstic or similar formations with large openings. Also, in the vicinity of large water bodies and streams Darcy's law may not be found entirely valid.

In hydrogeological interpretation, the hydraulic conductivity k has been also called the coefficient of permeability. At the U.S. Geological Survey the coefficient of permeability was originally defined as the flow of water at 60 °F in gallons per day through a medium having a cross-sectional area of 1 ft^2 under a hydraulic gradient of 1 ft, for laboratory conditions, and as the flow of water in gallons per day under a hydraulic gradient of 1 ft $mile^{-2}$ at field temperature.

In the saturated zone, the coefficient of transmissibility was also introduced. This equals the field coefficient of permeability (in metres per day) multiplied by the aquifer thickness M in metres:

$$T = kM \quad m^2\,day^{-1} \qquad 2.16$$

For example, when pumping tests give the transmissibility coefficient T as

$2.1 \cdot 10^{-3}\,m^2\,s^{-1}$ and the aquifer thickness is 40 metres, the coefficient of permeability, k, equals:

$$\frac{2.1 \cdot 10^{-3}}{40} = 5.25 \cdot 10^{-5} \qquad m\,s^{-1}$$

From the contours of the groundwater level the groundwater head was found to be 80 metres at a distance of 10 kilometres, the gradient is 0.008 and the flow velocity is

$$v = kI = 5.25 \cdot 10^{-5} \cdot 0.008 = 4.2 \cdot 10^{-7} \qquad m\,s^{-1}.$$

Taking the noncapillary porosity of the aquifer $p = 0.15$, the effective velocity is

$$v_e = \frac{4.2 \cdot 10^{-7}}{0.15} = 2.8 \cdot 10^{-6} \qquad m\,s^{-1}$$

This velocity would be observed by using dye or radioactive tracer.

Obviously by measuring v_e, k can be determined also:

$$k = \frac{pv_e L}{h} \quad \text{where} \quad \frac{L}{h} \quad \text{is the gradient.} \tag{2.17}$$

In the absence of field measurements k can be estimated using Fair and Hatch's formula

$$k = \frac{1}{5\left[\frac{(1-\alpha)^2}{\alpha^3}\left(\frac{\Theta}{100}\sum\frac{P}{d_m}\right)^2\right]} \tag{2.18}$$

where α is porosity, Θ is the sand factor varying from 6.0 for spherical grains to 7.7 for angular grains, P is the percentage of sand held between adjacent sieves and d_m is the geometric mean of rated sizes of adjacent sieves. A uniform system of sieves has to be applied.

Several instruments have been developed for measuring k in the laboratory. An example of constant-head permeameter is shown in Fig. 2.11. Here

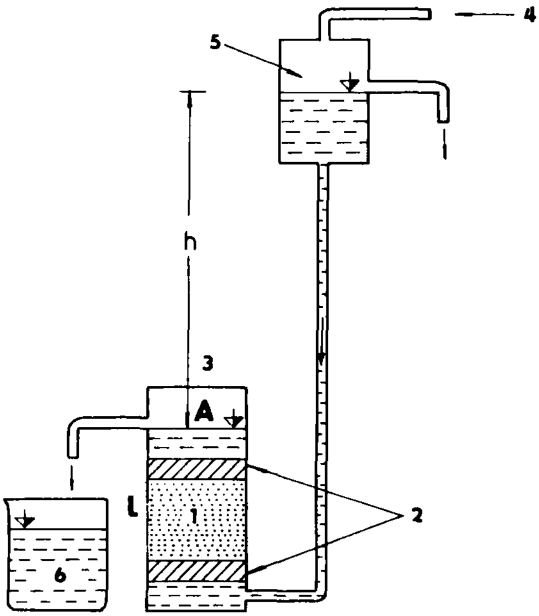
$$k = \frac{OL}{Ath}$$

where O is the flow volume at time t and the other quantities are as given in the figure.

It should be noted that the difference between values of k obtained in the laboratory and from field measurements can be rather high.

For unsaturated soil, the water movement can be described by an equation analogous to Darcy's law.

$$v = -k_k \operatorname{grad} \Psi \tag{2.19}$$



2.11 Permeameter. Sample – *1*, porous plate – *2*, horizontal area of sample – *3*, water supply – *4*, constant water level – *5*, volume O at time t – *6*.

where grad Ψ is the gradient of the soil water potential and k_k is the capillary conductivity. To determine the capillary conductivity, as one of many approximations Gardner's formula can be given [30]:

$$k_k = \frac{1}{(p_s/p_{sk})^n + 1} \tag{2.20}$$

where k_k is the coefficient of permeability: p_{sk} is the soil water tension, negative when $k_k = k/2$ (usually $1 \sim 5$ kPa); p_s is the soil water tension corresponding to k_k; and n is the coefficient for a given soil (for fine materials $n = 1 \sim 2$, for sands $n = 10$).

Also it was introduced

$$k_k = \frac{a}{\dfrac{p_s}{100} - 1000} \tag{2.21}$$

where p_s is soil water tension in Pa, and a is a constant for a given soil. The equation is valid for $p_s < 200$ Pa. A rational simplified solution of a saturated-nonsaturated zone system, based on the budgeting, is given in Chapter 6.

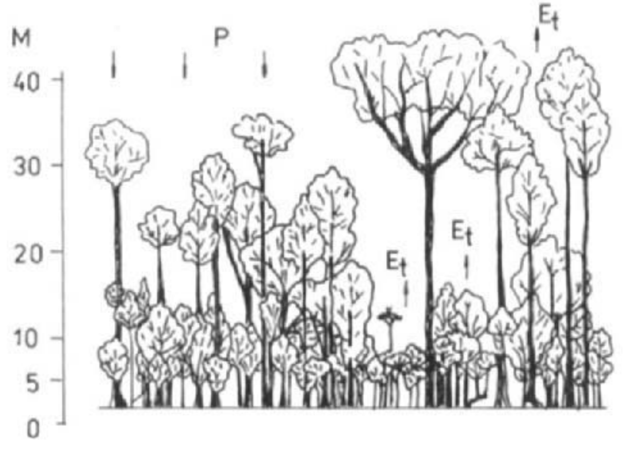
2.8.2 Evapotranspiration in the water balance approach

While all the basic components of the water balance equation, namely rainfall, groundwater outflow and surface runoff are subject to downward movement and under the dominant influence of the force of gravity, evapotranspiration is characterised by upward movement; the main driving force being the energy emitted by the sun. At first sight this process may appear to be of minor significance in hydrogeological studies. However, bearing in mind that the process is closely related to the atmosphere-plant-soil-groundwater system, sooner or later we discover how significantly the groundwater régime and its formation is affected by evapotranspiration. The main impact of evapotranspiration is in the domain of groundwater recharge, where the potential recharge is predominantly controlled by the combined effect of precipitation and evapotranspiration. The first controlling factor over groundwater recharge is applied when part of the rainfall falling on the surface of the vegetation is captured, and this intercepted water is partly released back to the atmosphere (Fig. 2.12). Then part of that water which reaches the ground flows away as surface runoff, and only the remaining water can infiltrate into the soil.

The next controlling factor over the amount of water to be recharged is the soil itself. The infiltration rate, highly variable during the first hour, is closely linked with the physical properties of each particular soil type. While above the ground the water can be lost to the atmosphere as evaporation, immediately after it enters the soil profile it becomes subject to soil evaporation and/or transpiration. From such a point of view, groundwater recharge appears to be rather an exceptional event in many regions. Nevertheless, in many groundwater models it is simulated as a continuous, throughout-the-year, process.

Not only groundwater is affected by the evapotranspirational process. In wet and dry regions of the tropics some plants depend primarily upon groundwater storage, which they tap through a well-adapted system of deep roots, and some species are able to consume 90% of their water requirements from stored groundwater in the zone of capillary rise.

Therefore in the water balance approach it is essential to consider all the components of the evapotranspirational process as great consumers of potential groundwater recharge.



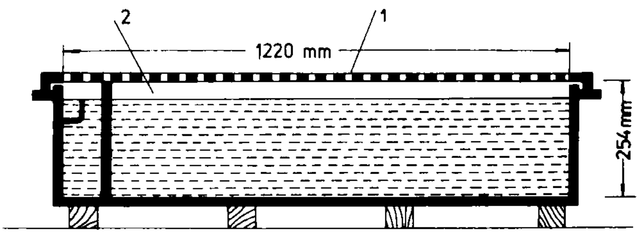
2.12 Interception loss varies in accordance with meteorological situation, season and exposure of the vegetation.

2.8.3 Evaporation from free water surface and from snow

The energy of about 2500 Joules is required to vaporize one gram of water. The energy supporting this process comes from solar radiation, and the sun is a dominant factor in the evapotranspirational process. The seasonal variation of the latter according to the latitude and the occurrence of meteorological phenomena such as cloud cover, wind speed, air temperature and air humidity, is closely linked with solar radiation. In the absence of direct radiation measurement, the availability of energy is often judged by air temperature.

The problem of assessing evaporation has long been a matter of concern to hydrometeorologists. Direct measurements using evaporation pans, various types of atmometers, and formulas based on an aerodynamic approach are often applied at present. However, empirical formulas are still widely used in the absence of accurate meteorological measurements.

For the direct measurement of evaporation from a free water surface, the U.S. Class A Pan commonly is used. It is a cylindrical tank, 1.22 m in diameter and 0.254 m deep, filled with water to a fixed level (Fig. 2.13). It is supported above the ground on a wooden platform of 100 · 50 mm joists and 100 · 100 mm bearers. As the purpose of the platform is to allow free circulation beneath the pan, the platform is of open slat construction with 50% open space. The water level in the pan must be measured accurately before and after water is added to the pan. A calibrated cup is used to add or remove water and the amount of water is recorded



2.13 Standard evaporation pan. Protective wire mesh – *1*, stilling pool – *2*.

on a form, with adjustments for rainfall. The water level must be maintained at the level of the indicator mounted inside the evaporation pan.

Use of this pan under tropical conditions indicated that under direct sunshine the metallic pan becomes much hotter than the water it contains, and thus the measurements are influenced seriously. Several modifications of the design have been proposed, but these would detract from standardization. Another problem in some regions is protection of the pan against animals and birds.

In the absence of direct measurements, various types of formula are applied.

Penman's formula has become widely used in the world and is used for the calculation of the potential evaporation E_0:

$$E_0 = \frac{p_0}{p}\frac{\Delta}{\gamma}\left[0.95\,R_A\left(a + b\frac{n}{N}\right) - \sigma T_K^4(0.56 - 0.079\sqrt{e_a})\left(0.1 + 0.9\frac{n}{N}\right) + \right.$$

$$\left. + 0.26(e_0 - e_d)(0.5 + 0.54U)\right]\left(\frac{p_0}{p}\frac{\Delta}{\gamma} + 1.00\right)^{-1} \qquad 2.22$$

The terms in the formula are expressed as follows:

E_0 – estimation of the evaporation from a free water surface for a given period, expressed in mm,
p_0 – mean atmospheric pressure, expressed in Pa, at sea level,
p – mean atmospheric pressure, expressed in Pa as a function of altitude for a given station,
Δ – rate of change with temperature of the saturation vapour pressure, expressed in Pa per degree °C,
γ – the psychrometric coefficient for the psychrometer with forced ventilation = 0.66,
0.95 – factor expressing the reduction in the incoming short wave radiation on the evaporating surface of water and corresponding to an albedo 0.05; for the evapotranspiration from plants and crops a corresponding albedo and factor are to be introduced,
R_A – short wave radiation received at the upper limit of the atmosphere, expressed in mm of evaporable water,
a and b – coefficients for the estimation of total radiation from the sunshine,
n – sunshine duration for the period considered, in hours,
N – sunshine duration as a possible maximum for a given location, in hours,
T_K^4 – black body radiation expressed in mm of evaporable water for prevailing air temperature,
e_a – saturation vapour pressure, expressed in Pa,
e_d – vapour pressure for the period under consideration, expressed in Pa,
U – mean wind speed at an elevation of 2 m for the given period, expressed in m s^{-1}.

This formula indicates how complicated is the whole process. To speed up the calculations, various types of step-by-step methods are available, including tables necessary for the calculus.

At some stations the Piche evaporimeter is used (Fig. 2.14) in which the water is evaporated on filter paper. On Wild's evaporimeter, the evaporating tank is connected to small scales.

As an example of empirical formulas, Linacre's method has been recommended for semi-arid regions (Schulze [31]). Only the air temperature record is required for the calculation:

$$E_0 = \frac{500T_m(100 - L) + 15(T - T_d)}{80 - T} \quad \text{mm day}^{-1} \qquad 2.23$$

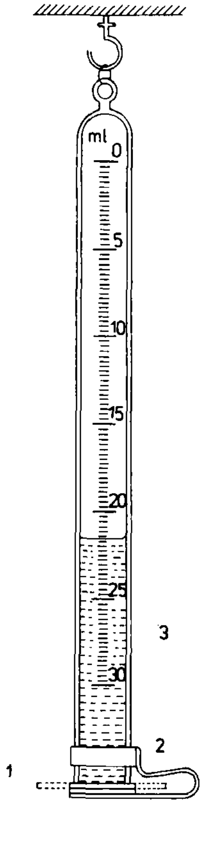
where T is mean temperature in °C for a given day; $T_m = T + 0.006A$ with A being elevation above sea level in metres; L is the latitude in degrees; and $T - T_d = 0.0023A + 0.37T + 0.53R + 0.35R_{an} = 10.9$, in which R is the mean daily range of temperature, and R_{an} is the difference between the mean temperature of the hottest and coldest months of the year, with both values given in °C.

A formula of this type should be tested at some observation site before it is widely applied in the region.

An even simpler formula was introduced by Gonzales and Gauga [32] for Central America:

$$E = 0.53t - 3183 \qquad 2.24$$

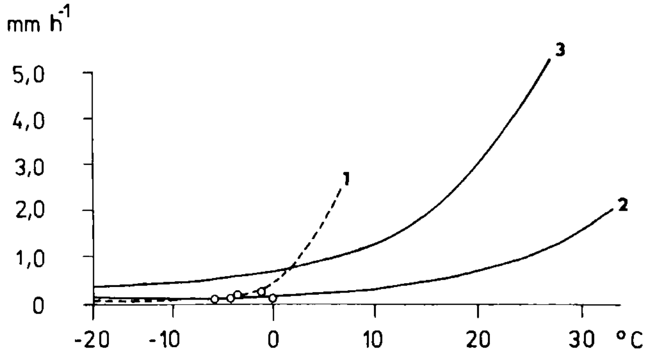
where E is the annual evaporation from a free water surface, in mm, and t is the sum of mean daily temperature exceeding 10 °C.



2.14 Piche evaporimeter. Filter paper – *1*, spring clip – *2*, distilled water – *3*.

As far as evaporation from a snow surface is concerned, the process has been studied jointly with the vapour condensation on the snow surface [33]. In addition to the amounts of water evaporated or condensed, the latent heat of vaporisation involved in the change of state from gas to liquid or vice versa is also involved. It has been found that the latent heat is capable of producing more than sevenfold decrease in the amount of water available for runoff.

Polyakov [34] found a relationship between snow evaporation and temperature, which can be used as a guide when comparing it with the evaporation from free water surface (Fig. 2.15).



2.15 Evaporation from snow after Polyakov – *1*, evaporation at field capacity after Kostin – *2*, and evaporation at soil saturation after Kostin – *3*.

In the evaporational process, the evaporation from intercepted water is included. From the point of view of groundwater recharge, water evaporated from the vegetation surface represents a great amount of water which is effectively lost, and which could otherwise contribute significantly to the groundwater resources. Thus at first, vegetation may act as a barrier against the potential recharge of groundwater. However, where there is no vegetation in the region and the rain falls on bare soil, as is typical in semi-arid and arid regions, it results in a high surface runoff and the development of erosional processes. Thus the establishment of a well balanced vegetation ecosystem is, from the point of view of groundwater hydrology requirements, a challenging task for future work towards effective groundwater management.

A simple method of measuring of evaporational loss through interception would appear to be the comparison of measurement of rainfall under the tree crown and on the open area close to the tree. Some difficulties during the winter period can be expected, and in fact the results of such a measurement are usually effective only for the growing season. Here, however, the effect of horizontal rainfall must be always taken into account. Such types of experiment provide widely varying results.

In tropical regions, the interception was analysed by Jackson [35] who concluded that about 20–30% precipitation is lost by evaporation from interception in the jungle, 20% in the savanna and 15–20% in the tropical croplands. Experiments in humid regions (Němec at al. [36]) indicate that a mixed forest of oak and spruce evaporates 20% of annual rainfall, pure fir forest 55% and fir-spruce mixture 49%. A variety of interception values from year to year and from stand to stand can be expected.

Provided the measurement of transpiration *in situ* is available, the water balance approach can be applied. As an example data from the experimental station in Czechoslovakia can be given for 1982:

Rainfall: 820.76 mm,
surface runoff from spruce-oak mixture: nil,
transpiration measured for spruce-oak mixture: 206.6 mm,
evaporation of intercepted water: 483.64 mm.

These results indicated that evaporation from intercepted water was about 234% of the transpired water. This value seems to be relevant to the conclusion reached by Rutter [37], who in England estimated that four times more water was lost from intercepted water than from transpiration, and of Monteith [40] who measured five times more in a coniferous forest.

These values only indicate how important is the rôle played by interception in the formation of groundwater resources.

2.8.4 Evaporation from soil

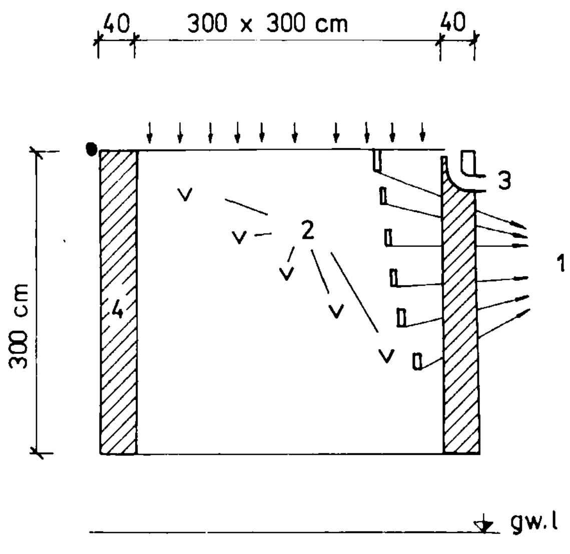
Unlike evaporation from a free water surface, this process is affected by the nature of the surface from which the evaporation takes place. Also, when evaporating from soil, the water molecules have to overcome greater resistance. Thus in addition to the meteorological phenomena affecting the evaporation from free water surface, the soil structure itself becomes an additional factor together with soil moisture content. It has been assumed that the evaporation loss from soil practically ceases when the soil moisture content is close to the wilting point. Some authors believe that evaporation from a soil surface will continue as long as the shallow surface layer (10 – 20 cm) remains moist. Veihmayer and Hendrickson [39] have found that the evaporation rate remains constant until the wilting point is reached.

Kostin [38] presented curves of evaporation, from fully saturated soil and soil at field capacity, as dependent on the air temperature. Even if the relationsips are simplified (Fig. 2.15), all curves provide some guidance on the soil evaporation process. The most convenient field instrument for assessing the soil evaporation is the lysimeter. Weighed lysimeters have been widely used for the measurement of soil evaporation and also of evaporation by crops. Typically it consists of two cylinders and a water collecting well. The inner cylinder has a movable perforated bottom bolted at three points. Usually a measuring device for the measurement of surface runoff from the lysimeter is attached. The water flowing through the lysimeter is also collected. The water balance and evaporation are measured by weighing the soil-filled tank.

The disadvantage of the weighing lysimeter is the small surface area, which may not be sufficiently representative. Extension of the surface area results in difficulties in handling and maintenance of the device.

Provided there is an instrument available for measuring soil moisture at different soil horizons, it can be more convenient to define the soil moisture at several horizons. Vertical drains are installed at various depths, which collect excess water from non-capillary pores and thus measure the contribution to the groundwater recharge. The surface runoff can be measured by a V-notch weir, or in the water collecting tank attached to the lysimeter surface. The soil evaporation from bare soil is calculated by the water budgeting method, which means that the precipitation has to be measured close to the lysimeter. An example of such a lysimeter is in Fig. 2.16.

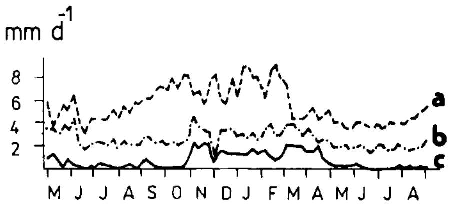
For the purpose of groundwater resources assessment, a lysimeter of this type should be freely connected with the groundwater level through the zone of capillary rise.



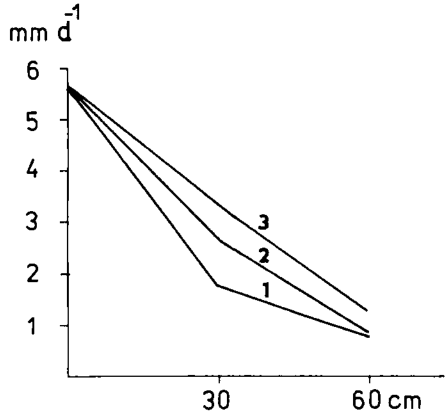
2.16 In situ lysimeter, Nedamov, Czechoslovakia. Capillary water monitoring – *1*, non-capillary water measurement – *2*, surface runoff measurement – *3*, concrete insulated walls – *4*.

In specialised studies concerned with the relationship between soil evaporation and groundwater table, experimental work is frequently involved, particularly when a limited amount of data are available. Already in 1932 White [41] proved that when the zone of capillary rise approaches the zone of aeration, a significant loss occurs due to the evaporation from groundwater sources. He concluded that with a groundwater level 1 metre below the surface, 7% of the pan evaporation can be expected; with the level at 50 cm it is 15%; and with the level at 25 cm it is 30%.

From Hellwig's observation [42] a pattern of decreasing evaporation rate with the decrease of the groundwater level can be traced (Fig. 2.17). From such observations a generalised relationship between water table depth and soil evaporation (Fig. 2.18) can be obtained.



2.17 Fluctuation of evaporation from sand as a factor of groundwater depth in a medium sand mixture. Water table at the surface – *a*, 30 cm below sand surface – *b*, 60 cm below sand surface – *c*. After Hellwig's experiments in Namibia.



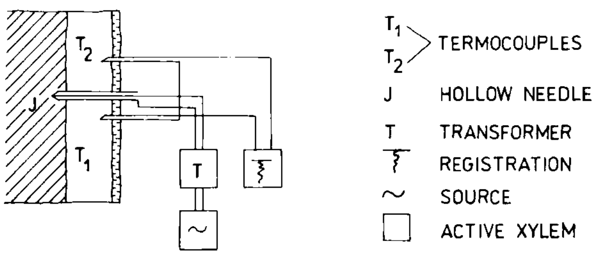
2.18 Evaporation from sand as a function of the depth of groundwater table. Coarse sand – *1*, medium sand – *2*, fine sand – *3*. After Hellwig's experiments in Namibia.

2.8.5 Transpiration

This phenomenon is the factor most difficult to measure. In literature many methods have been described which are in fact concerned with the measurement of evapotranspiration, and in some cases the interception is also included.

Actual transpiration is related only to the water movement through plants, or in other words, through the stem of a plant, and it is defined as the water exchange between soil and atmosphere through the root system.

Swanson [43] described a flow meter for estimating the rate of sap ascent in trees. The method (Fig. 2.19) is based on the measurement of heat conduction through



2.19 Swanson's sap flow meter.

the flowing sap. Two thermocouples are installed in the active xylem in the plant. A hollow needle, through which the thermal source is injected into the plant, is placed between them. The temperature difference between the thermocouples T_1 and T_2 is continually measured and registered. Variations in the temperature difference thus characterise variations in the sap stream velocity, or it could be said, variations in the transpiration.

The heating is continuous, but very low amounts are applied to avoid damage to the plant structure. About 250 mV is recommended for a needle 3.5 cm long.

During typical meteorological situations the instrument is calibrated by generating single heat pulses, higher but of short duration. Impulses of about 30 W and duration of 1 sec have been successfully applied. The velocity calculation is based on the assumption of no heat losses, radial heat propagation and uniform flow. The velocity is calculated from the response of the temperature difference to an impulse occurring at t = 0:

$$\Delta T = \frac{Q}{\varrho c}\frac{1}{4}\pi kt \exp\left[-(x - Vt)^2 4kt\right] \qquad 2.25$$

where T is the temperature difference, K; Q quantity of heat applied per unit length of needle, J m^{-1}; ϱ density, kg m^{-3}; c heat capacity J kg^{-1} K^{-1}; k diffusivity, m^2 s^{-1}; t time, s; x distance from the heat source, m; V velocity of flow, m s^{-1}. Thus from two equations set up for two points with different values of x, we obtain the temperature pulse or sap stream velocity

$$V = \frac{x_1 - x_2}{2t_0} \quad \mathrm{m\,s^{-1}}; \qquad x_1 \neq x_2, \qquad 2.26$$

where t_0 is the time interval (s) between the impulse and the time at which the temperature at both measured points is the same, and x_1 and x_2 are the distances of the thermocouples from the thermal source, m.

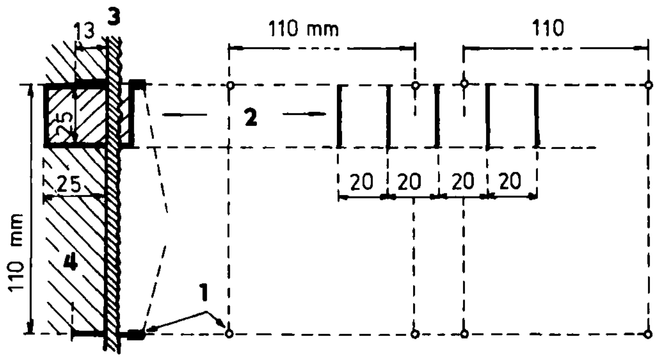
Usually the fluctuation of the net radiation, air temperature, rate and duration of rainfall, wind speed, humidity and soil moisture are also registered, which provides information for formulating a relationship of regional validity.

Using such a type of instrument (Balek and Pavlík [26]), a multiple regression equation was set up for the calculation of sap stream velocity of spruce (*Picea abies*):

$$V = 85.53 \exp(0.003R + 0.0265T - 0.008W - 0.0015S - 0.34D) \qquad \mathrm{cm\,day^{-1}},$$

where V is sap stream velocity, cm day^{-1}; T mean daily temperature, in °C; R net radiation, in cal cm^{-2} day^{-1}; W mean daily wind velocity, in km h^{-1}, S daily rainfall, in mm, and D rainfall duration, in hours.

To determine total flux through the active xylem, an additional assessment of the xylem cross-section has to be done. Therefore the method was modified (Balek et al. [44]). The measuring system consists of a power generator, a series of thermocouples, electrode and a recording system (Fig. 2.20). Power is supplied to the tree trunk through five stainless steel electrodes installed in the hydroactive xylem. The temperature difference between heated and nonheated parts of the measured trunk element is measured by eight thermocouples. Two of them are installed at the upper vertices of the central electrodes, two at their lower vertices and four on either side of the measuring device. The installation is thermally insulated and shaded from direct sunshine.



2.20 Measuring instrument for direct measurement of transpiration, after Čermák et al. [45]. Thermocouples – *1*, steel electrodes – *2*, tree bark – *3*, active xylem – *4*.

The basic equation for the direct calculation of the transpirational flux of sap through a single tree is

$$Q_{wt}^{rec} = \frac{P}{c_w \Delta T}\left[\frac{O_{bk}}{d}(n-1)\right] \quad \mathrm{l\,s^{-1}} \qquad 2.27$$

where P is the power input, W; ΔT is the temperature difference between the heated and nonheated parts of the tree, K; d is the distance of the electrodes, cm; n is the number of the electrodes; O_{bk} is the circumference of the tree without bark at the measured element, cm; and c_w is the specific heat of water, 4186.8 J K^{-1}.

The actual discharge deviates slightly from the measured one because of the thermal loss. Therefore we get

$$Q_{wt} = Q_{wt}^{rec} - Q_{wt}^{fic} \qquad 2.28$$

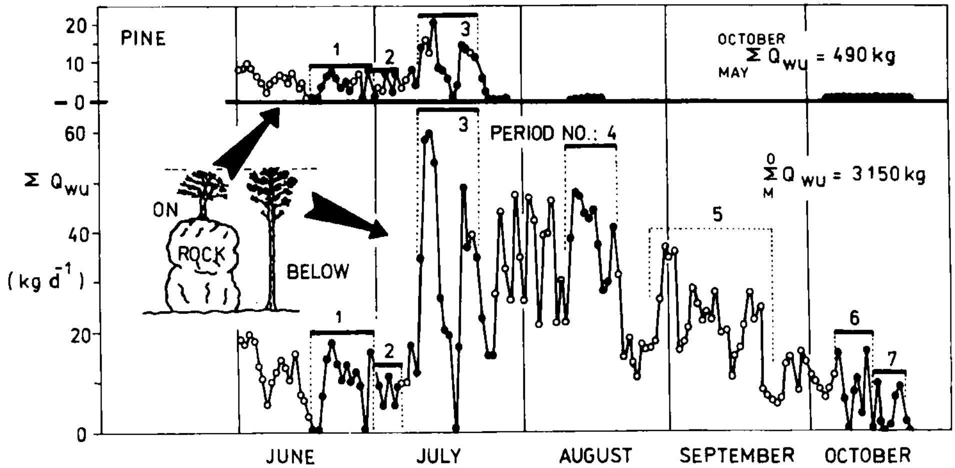
where Q_{wt} is the actual discharge through a single tree, Q_{wt}^{rec} is the recorded discharge and Q_{wt}^{fic} is the so called fictive discharge, recorded when actual discharge can be expected to equal zero, usually before sunrise.

Areal representativeness can be calculated:

$$Q_{ws} = Q_{wt}\frac{k_s}{k_t} \quad \mathrm{l\,s^{-1}} \qquad 2.29$$

where k_s is the size of the basal area of all trees in the representative plot A, and k_t is the mean basal area of the tree used in the measurements (both values in m^2).

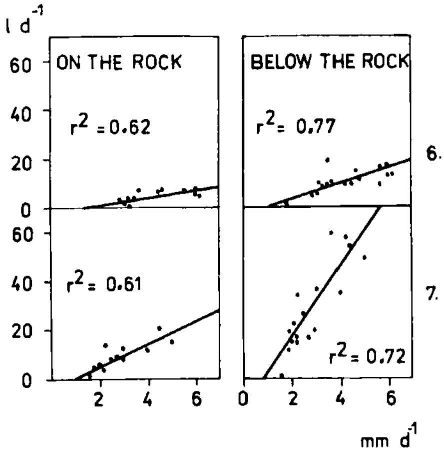
Using the above device a continuous record of the transpiration can be obtained, as in Fig. 2.21.



2.21 Transpiration from sandy loam overlying sandstone formations, and from shallow soil profile on sandstone rock; Nedamov, Czechoslovakia.

Using the method for groundwater recharge assessment in sandstone formations and temperate climate, the formula for the transpirational loss was obtained for *Pinus silvetris L.* on rock with a shallow soil profile less than 70 cm deep (the soil actually fills the rock fissures):

$$Q = (0.270V - 1.2)\frac{\Delta H + \gamma_A}{\Delta + \gamma} - 0.280V + 1.09 \ \ \mathrm{l\,day^{-1}}$$



2.22 Relationship between the transpiration rate and potential evaporation as dependent on the seasonal development of soil water content; June/July, Nedamov, Czechoslovakia.

Similarly for pine growing beside the rock on sandy loam with the unsaturated zone depth about 200 cm:

$$Q = (0.0376V + 1.97)\frac{\Delta H + \gamma_A}{\Delta + \gamma} - 0.0297V - 2.11 \qquad \mathrm{l\,day^{-1}}$$

In both formulae the potential evaporation, calculated by Penman's formula, served as a guide to the relationships as obtained in Fig. 2.22. Thus in the formulae ΔH is the heat budget term, γ_A is the aerodynamic term, Δ is the variation of saturation water vapour pressure with temperature, γ is a psychrometric constant from Penman's equation, and V is the soil moisture content (mm) in the soil profile.

The first formula was found to be valid for the tree on the rock when the potential evaporation E_0:

$$E_0 > \frac{0.280V - 1.09}{0.270 - 1.2} \qquad \mathrm{mm}$$

and for the tree on sandy loam

$$E_0 > \frac{0.0297V + 2.11}{0.0376V + 1.97} \qquad \mathrm{mm}$$

The limits of validity actually represent temperature boundaries between growing and non-growing seasons.

Fig. 2.22 indicates a close relationship between potential evaporation and transpiration. However, the changing gradient of the lines indicates the additional effect of the soil water content and its seasonal effect on the transpirational process.

Through repeated measurements, some idea can be obtained of the amount of water which is lost to the atmosphere. Combining the results with the outflow on the area involved, we thus obtain information on groundwater recharge pattern in the area (Tab. 2.3).

Knowing that for coniferous trees in studied area the interception rate is about 70%, we find for the water balance values in the table following groundwater recharge G:

$$G = 374.3 - 0.7 \cdot 374.3 - 4.36 = 107.9\ \mathrm{mm}$$

for sandstone rock formations and

$$G = 374.3 - 0.7 \cdot 374.3 - 47.15 = 65.14\ \mathrm{mm}$$

for sandy loams. The assessment is based on the values given in Tab. 2.4 and supposes that surface runoff in the forest is equal to zero.

The direct method cannot be applied on vegetation with thin stalks. Instead a lysimetric method as described for bare soil can be used. The type of vegetation to be studied is usually planted on the surface of the lysimeter. On standard size lysimeters, grass and various types of crops have been planted; at the Soviet experimental station of Walday lysimeters of giant size were planted with coniferous trees.

Tab. 2.3 Monthly values of factors which influence the transpiration and amount of water transpired by trees on sandstone rocks and on sandy loams

Month	Rainfall	Radiation	Temperature	Potential evaporation	Regional transpiration rock	sandy loam
	mm	J cm^{-2}	°C	mm	mm	mm
April	30.0	25.126	7.9	89.0	0.537	1.898
May	20.5	64.925	14.6	126.5	2.039	12.000
June	114.3	65.110	19.4	136.0	0.986	44.669
July	53.4	46.512	16.2	96.0	0.613	9.284
August	44.5	45.033	17.0	96.4	0.062	11.070
September	98.4	29.531	13.6	59.7	0.062	6.500
October	13.2	21.294	7.1	76.0	0.062	1.732
Total	374.3	288.531	13.6	624.2	4.361	47.153

2.8.6 Evapotranspiration

Regional groundwater studies, calculations of the water balance of catchments, and studies of the impact of land use on the groundwater régime, are concerned with the water loss from a whole region. This type of regional study, will include all types of land use, selected crops, urban areas, roads, etc. In effect, the task is to calculate the combined effect of all types of evapotranspiration, including the evaporation from both free water surface (such as ponds, lakes, streams) and intercepted water, the evaporation from soils, and the transpiration by plants. Often we can only apply an aproximate solution. As mentioned when discussing Penman's method for the calculation of potential evaporation, the factor expressing the reduction in the incoming short wave radiation on the evaporating surface can be replaced by another factor corresponding with the albedo (Tab. 2.4).

However, in the absence of data suitable for this type of approach, various types of approximate solutions have been applied. Among many rational formula Turc's has been widely used:

$$E_t = \frac{P}{\left[0.9 + \left(\frac{P}{L}\right)^2\right]^{1/2}} \quad \text{mm year}^{-1} \qquad 2.30$$

where P is the annual precipitation, in mm; $L = 300 + 25T + 0.05T^3$; T is the mean annual temperature, in °C; and E_t is the total evapotranspiration, in mm year^{-1}.

In the U.S.A. Thornthwaite's formula [48] has been widely used by agriculturalists:

$$E_t = 1.6\left(\frac{10T}{I}\right)^m \quad \text{cm month}^{-1} \qquad 2.31$$

Tab. 2.4 Albedo of various types of surfaces

Surface	Albedo %
Water, solar altitude 40°	2 – 5
Water, solar altitude 5 – 30°	6 – 40
Grass	15 – 30
dry dark soil	14
wet dark soil	8
sand	15 – 25
forest	3 – 10
fresh snow	80
ice, old snow	50 – 70

where T is the mean monthly air temperature, in °C; I is the annual heat index, and m is a cubic function empirically determined as $6.75 \cdot 10^{-7}I^3 - 7.71 \cdot 10^{-5}I^2 + 1.792 \cdot 10^{-2}I + 0.49239$ and

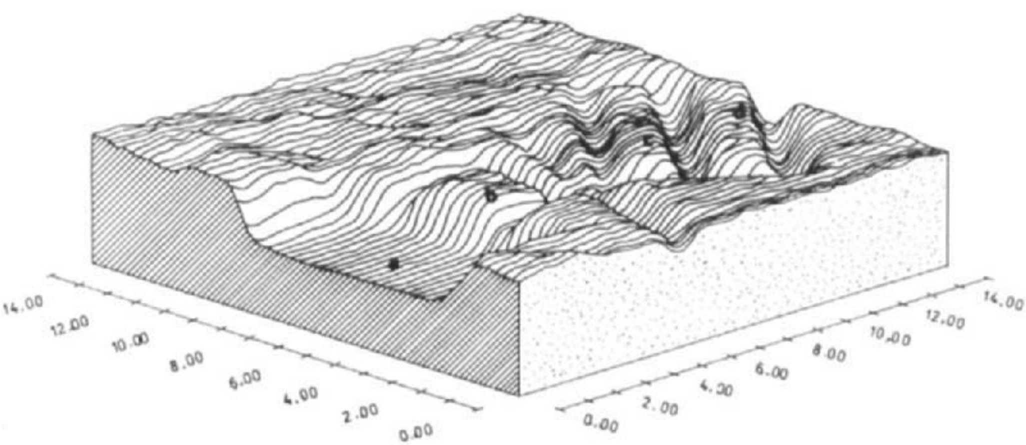
$$I = \sum_{i=1}^{n}\left(\frac{T}{5}\right)^{1.514} \qquad 2.32$$

Several attempts have been made to improve the validity of the formula by introducing other factors such as humidity, or other readily available climatic data.

2.8.7 Areal representation of the transpirational measurements by remote sensing

The results of transpirational measurements made as described in the preceding sections or measured by using lysimeters have a limited areal validity and it may be difficult to apply them regionally. In order to extend the measurements beyond the measured areas, remote sensing methods can be a convenient tool. It is, in fact, very difficult to apply remote sensing methods to hydrogeological studies if the area studied is covered by vegetation. Even the evaluation of surface soil moisture by remote sensing methods has been found difficult for any other than bare soils. Thus the groundwater régime can manifest itself only indirectly by remote sensing methods, as in this case through the areal measurement of the transpirational loss.

Earlier measurements on vegetation proved that plants reflect more energy in the infrared than in the visible part of the spectrum (Heller [46]). For some coniferous trees this ratio was 4 : 1 at the peak of the growing season and dropped after the growing season was over. It has been found that the reflectivity at the wavelength 0.8 – 1.1 µm (near infrared) is higher for leaves with a high content of mesophyllous cells and therefore the water content in the cell walls plays an important role. Therefore it is assumed that measurements of the rate of plant water transport can be based upon measurement of the energy reflected from the plant surface.

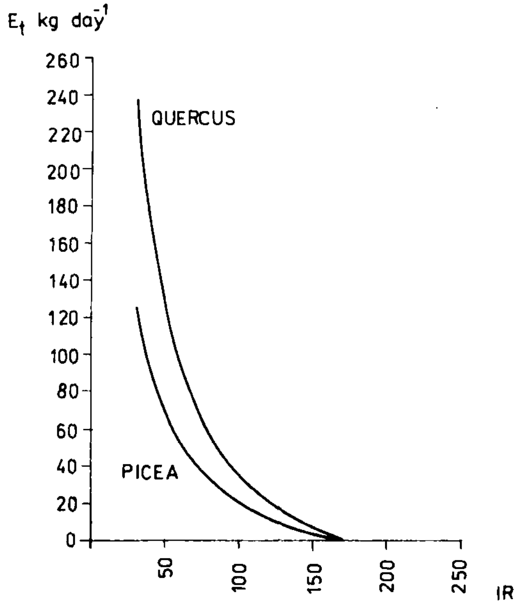


2.23 Three-dimensional image of optical density fluctuation in the monitored plot, band 5, July 1982, Nedamov, Czechoslovakia. Roof of the lysimeter house – *a*, grass in the lysimeter – *b*, bare soil – *c*, grass – *d*.

A low flying aircraft or a model aeroplane can be employed for this purpose, because then the pixels on the multispectral image represent small areas less than $20 \cdot 20\ cm^2$. Thus the crowns of the measured plants can be covered by a network of a number of pixels which is sufficient for statistical evaluation. Using infrared film the optical density of the bands 4 (0.5 – 0.6 µm), 6 (0.6 – 0.7 µm) and 7 (0.8 – 1.1 µm) of each pixel is determined, and then the matrix of all values is statistically evaluated. The optical density is measured in the range of 0 – 255 grey levels, with, in general, higher values of the optical density indicating lower reflectivity of the object.

The photographs selected after the visual evaluation are digitalised on a drum scanner, and the matrix of the optical densities is stored on magnetic tape.

The magnetic tape is processed by a computer and selected objects such as directly measured plants, lysimeters, swampy areas, bare soil etc., are analysed. A three-dimensional image serves for orientation. Fig. 2.23 shows an example of a three-dimensional image of the optical density in band 7 for a lysimeter and experimental plots for the measurement of surface runoff. The bright colour of the roof of the house attached to the lysimeter, visible on the picture, has stable optical properties and can be used for orientation and as a standard. The variability of the optical density of the lysimeter surface is rather low, which indicates a uniform transpiration. The optical density of the surrounding grass, which has not been cut, is more variable, and the slightly higher values of optical density indicate a lower transpiration from the lysimeter than from the surrounding meadow.



2.24 Relationship between optical density, band 7, and daily transpiration rate from oak and spruce.

On the surface runoff plots the optical density is higher for the bare soil than for the grass, which indicates higher water loss from the grass than from the soil. After analysing the three-dimensional image, mean values of the optical density are calculated for selected plots. A combination of ground measurements taken at the same time as the remote sensing measurement, and of the optical densities serves for further statistical analysis (Balek et al. [47]).

In an experimental area with spruce, the following relationship was found (Fig. 2.24):

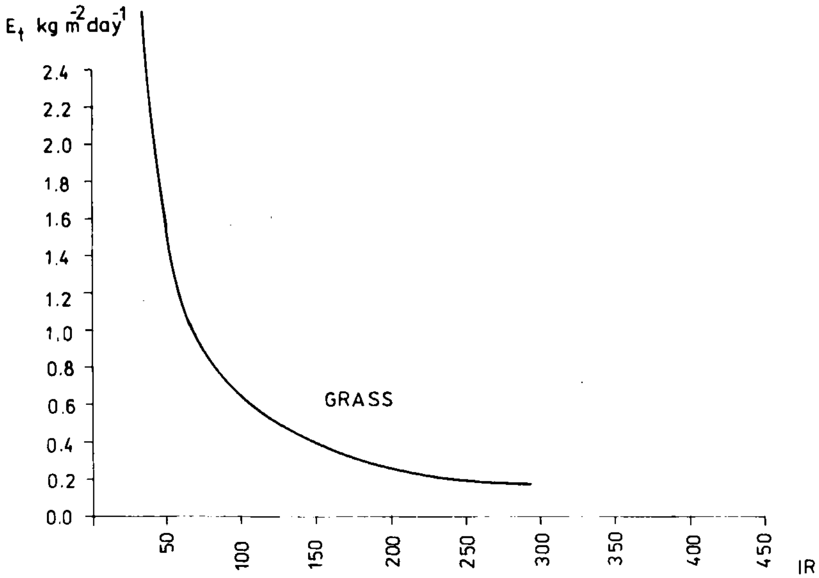
$$E_t^s = -27.3 + \frac{4654}{IR} \quad \text{litres day}^{-1} \qquad 2.33$$

where E_t^s is the daily transpiration rate in litres day^{-1}, and IR is the optical density in band 7. A similar relationship was found for oak:

$$E_t^o = -52.8 + \frac{8806}{IR} \quad \text{litres day}^{-1} \qquad 2.34$$

Finally for grass (Fig. 2.25):

$$E_t^g = \frac{1}{0.08IR - 0.024} \quad \text{litres m}^{-2}\,\text{day}^{-1} \qquad 2.35$$



2.25 Relationship between optical density, band 7, and daily transpiration of grass.

This type of relationship can be applied analogously to areas or regions with a similar type of vegetational cover, for which the mean value of the optical density in band 7 can be determined by aeroplane, satellite, etc. Because *in situ* observations can be made for combinations of various types of species, including the soil surface, a regional assessment of the water loss component in the water balance equation can be analysed using such remote sensing measurements. It should be noted, however, that for a more accurate assessment a special relationship should be developed, which is based upon *in situ* measurements in the region (Balek et al. 1986).

2.9 RÔLE OF METEOROLOGY AND CLIMATOLOGY

Meteorology is defined as the science concerned with the physical states and chemical processes of the atmosphere. Climatology is the study of long-term fluctuations in meteorological phenomena and the general atmospheric conditions of a region. Hydrometeorology is concerned with that part of the hydrologic cycle which occurs in the atmosphere.

Obviously there is a close link between meteorology, climatology and hydrometeorology, on the one hand, and hydrology and hydrogeology on the other. Meteorological and hydrometeorological data are required for many types of groundwater resources projects and evaporational and transpirational assessment; climatological data are needed for the assessment of groundwater resources and their development, for water balance calculations, and for groundwater exploitation strategies.

Of the many meteorological and climatological elements some are of primary importance: namely precipitation, radiation, air temperature, air humidity, wind speed and air pressure.

Climatology, concerned with the long-term fluctuation of meteorological phenomena and the general atmospheric conditions of a regional character, provides basic information for almost any hydrogeological study. The formation of groundwater resources, the fluctuation of groundwater recharge and any interaction of surface streams and groundwater resources are influenced by the climatological features. Also regionalisation is closely linked with the climatology. Historical climatology, periods of pluvials and interpluvials and similar phenomena are sometimes utilised when the formation of groundwater in the past is analysed.

Meteorological data, which provide information on the state of the atmosphere, are often used for the operational purpose of groundwater resources exploitation.

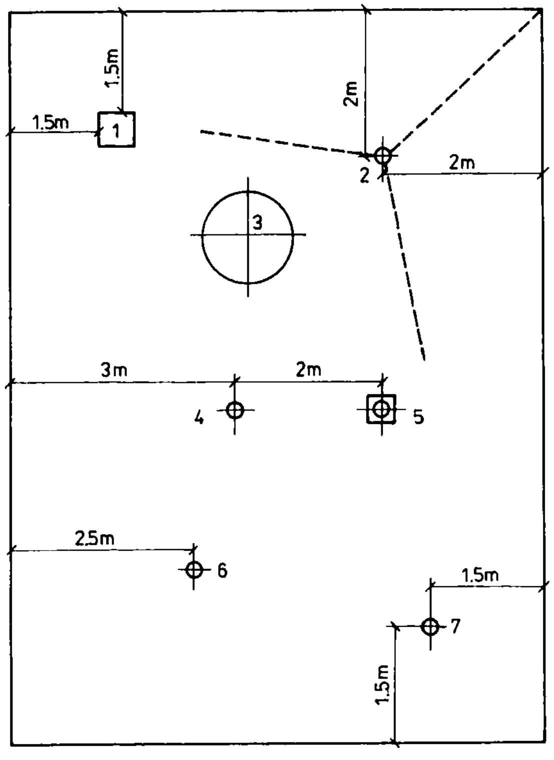
A hydrologist working on a hydrogeological project may discover that the available meteorological data are not adequate for a given purpose. Then one or more temporary observing meteorological sites have to be established at certain points. At these sites the precipitation, air and soil temperature, humidity, wind, sunshine, radiation, atmospheric pressure and evaporation will normally be measured.

It is important to select suitable sites for the exposure of instruments so that the data are representative for a given purpose. However, the location selected from the hydrological point of view does not always fulfill standard meteorological requirements. Nevertheless, the instruments should be freely exposed under all circumstances.

An ideal site would be a level area of ground approximately 9 by 6 metres, covered with short grass. Sites close to roads, especially bitumen roads, should be avoided because of heat and dust. The enclosure must be kept clean and no objects other than instruments should be housed inside. Buildings and constructions which would deflect or obstruct the natural flow of air must not be located near the fence. An example of a standard layout of meteorological instruments is given in Fig. 2.26.

A wide selection of standard instruments can be found in specialised catalogues. Working methods have been published in the WMO Guides. Automatic weather stations with data acquisition systems are increasingly used. Here the data are stored on magnetic tapes or memory modules, and can be directly processed on computers in an off-line or on-line system. New types of instruments with an electric signal are required for data acquisition stations. Nevertheless, it is recommended that an independent set of standard instruments should be installed in addition, which can be used for calibration and in the event of break-down of the automatic system.

It should be noted that often the meteorological data may contribute significantly to the analysis of groundwater record, and to the explanation of sudden changes in the groundwater levels. For instance, it is well known that an increase in atmospheric pressure may cause a decrease in the water level of boreholes penetrating confined aquifers. The vacuum effect of wind blowing across the top of a borehole, which causes a quick rise of groundwater level and other relationship, may be discovered also in the source of data analysis.



2.26 Meteorological observation site. Screen – *1*, wind mast – *2*, evaporation tank – *3*, rain gauge – *4*, rainfall recorder – *5*, sunshine recorder – *6*, earth thermometers – *7*.

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3 ANALYSIS AND SYNTHESIS OF THE WATER BALANCE COMPONENTS

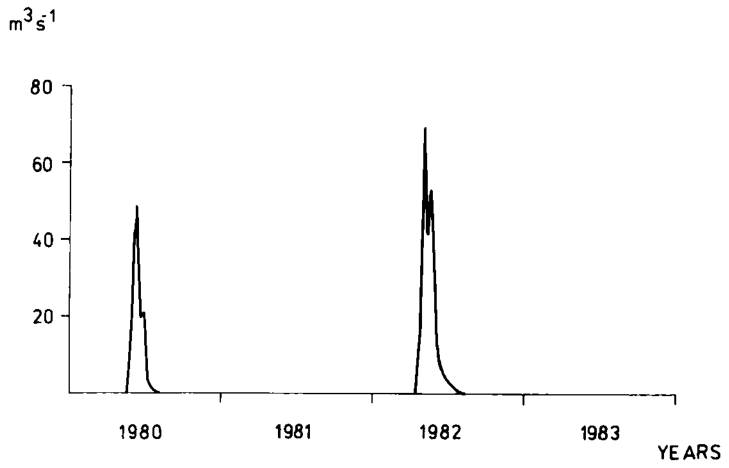
3.1 GROUNDWATER-STREAM NETWORK INTERACTION

River hydrographs contain essential information about groundwater resources. In general, any hydrograph is the result of a complex process of interaction between the surface runoff, subsurface flow and groundwater flow. The baseflow components of hydrographs are indicators of the natural replenishment of groundwater resources. This is particularly true of hydrographs with a long-term fluctuation of the river régimes. In some years, the formation of a baseflow may be due to exceptional recharge in the previous year, in another ones groundwater components of hydrographs correspond with groundwater sources which may have been formed, many years, decades and perhaps millenia ago.

By analysing annual hydrographs, three different patterns can be discerned:

1. hydrographs of ephemeral streams,
2. hydrographs of intermittent streams,
3. hydrographs of perennial streams.

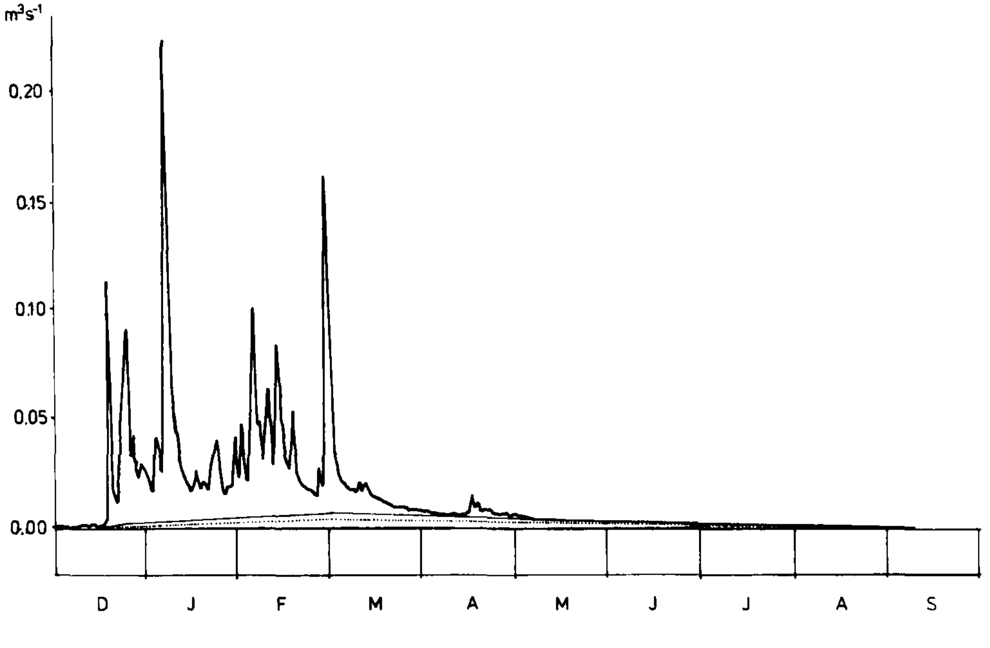
In hydrographs of ephemeral streams, typical for wadis (Fig. 3.1), we find that groundwater resources are practically non-existent. However, if the groundwater



3.1 Ephemeral régime, wadi type. Some flood discharge occurs occasionally from storm rainfall.

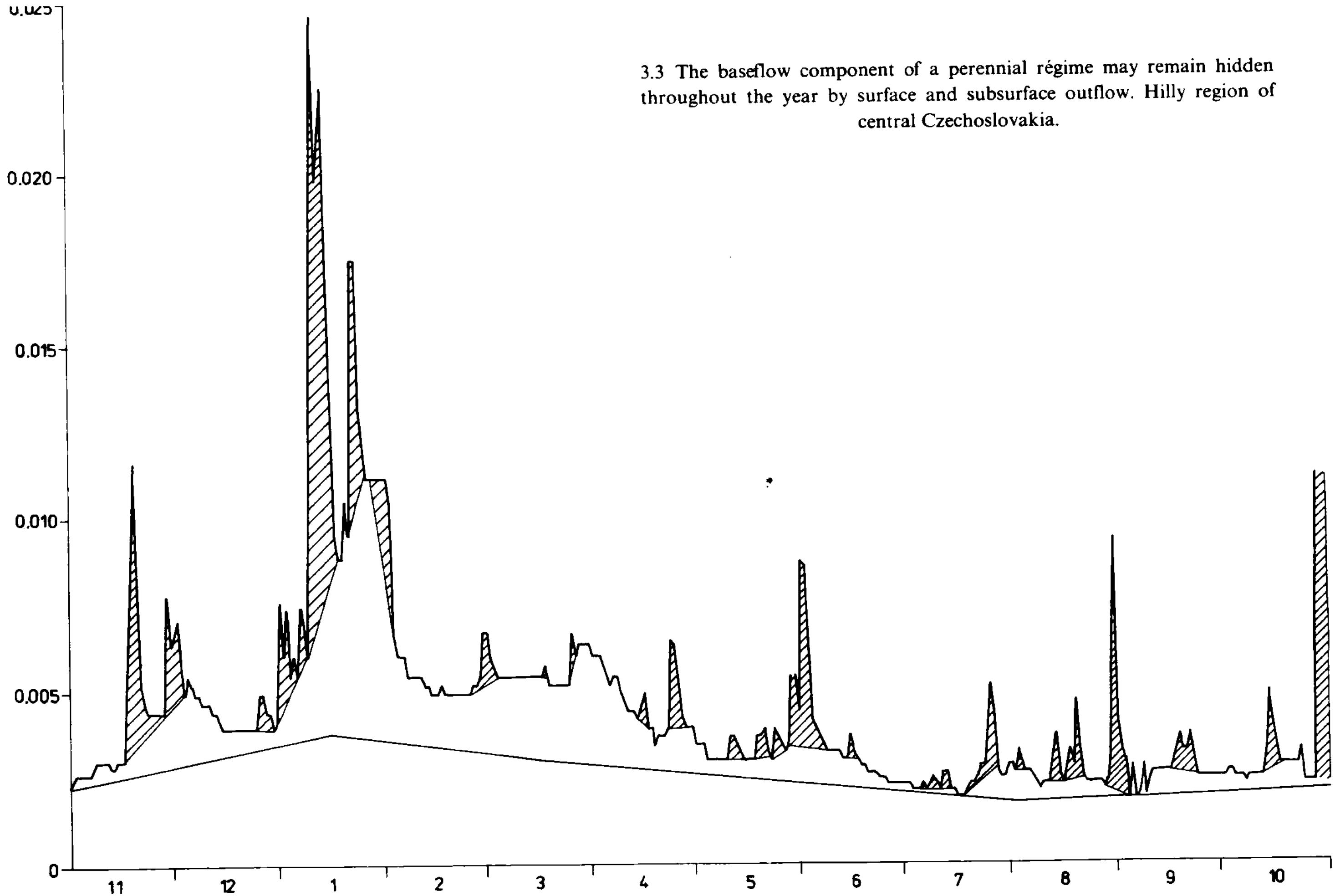
level fluctuates in the basin, some groundwater recharge may occur either from local rainfall or from historically or regionally exogenous resources. Fluctuations of the groundwater level may be caused by subsurface groundwater outflow, or by evapotranspiration from deep-rooted vegetation.

In the basins of rivers which flow into extremely arid regions and disappear there, due to intensive evaporation and/or seepage (as occurs for instance, with the Fafan and Gerer rivers in Ethiopia), some groundwater resources may be found in suitable locations below the river bottom, for some part of or throughout the year. These sources are often consumed by evaporation and phreatophytes.



3.2 Intermittent régime, headwater stream in the Zambezi basin. Groundwater outflow . . . . . , subsurface flow ——. Discharge diminishes at the end of the dry season.

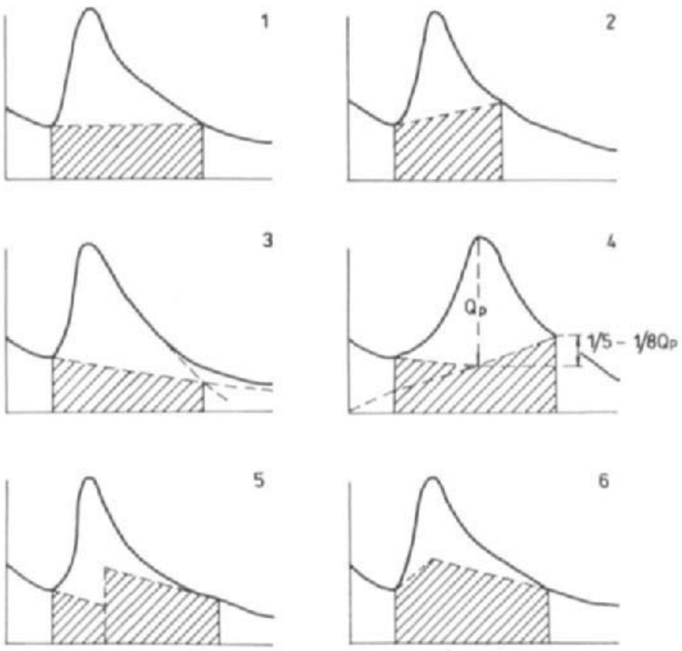
From the point of view of locating groundwater resources, intermittent streams are more promising (Fig. 3.2). In such cases, the shallow aquifers are usually replenishable only within limits, and the water in deep aquifers, even when it is in contact with the root system, may not be fully consumed for some years. In some hydrographs the river régime manifests itself every year, but in others it happens only occasionally. The groundwater fluctuation in wells and boreholes can be viewed as an indicator of whether the aquifers have been sufficiently replenished, and whether the streams will behave as perennial in that particular year.



3.3 The baseflow component of a perennial régime may remain hidden throughout the year by surface and subsurface outflow. Hilly region of central Czechoslovakia.

The search for baseflow components in hydrographs of perennial streams is usually concerned with the dry periods. The so-called "fair weather" parts of the hydrographs are those most convenient for the baseflow separation, particularly in regions with a pronounced dry season. In basins which have precipitation every month, separation is more difficult, because the groundwater component may be covered by the subsurface flow throughout the year, or at least for prolonged periods (Fig. 3.3).

Two basic methods can be employed for the separation of hydrograph components: the analytical approach when the observed hydrograph is separated into two or more partial hydrographs, and the synthetic approach when the hydrograph is synthetised from several partial hydrographs and the result is compared with the observed hydrograph.

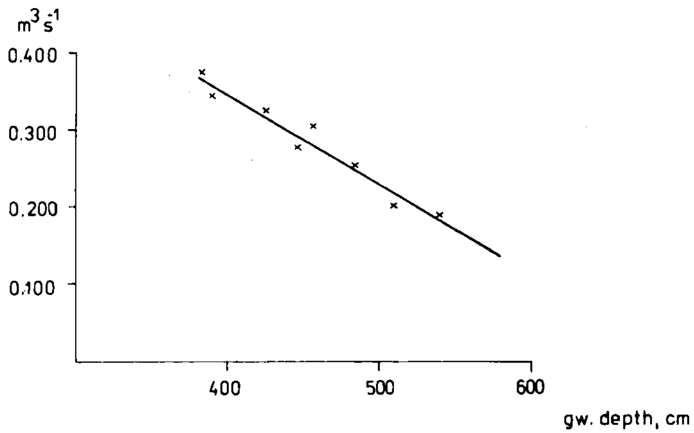


3.4 Simple separation methods.

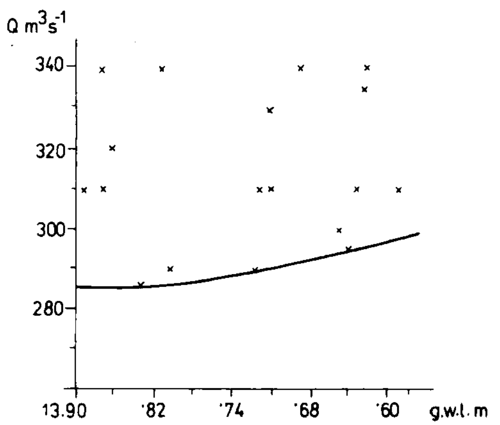
The evolution of separation methods can be traced back in time for many years. Such attempts were originally intended to separate the flood wave as a product of the so-called "effective rainfall". It is noticeable that, at that time, the contribution of rainfall to groundwater recharge was considered as somehow ineffective.

Examples of the separation methods typical of that period, and shown in Fig. 3.4, are self-explanatory. More methods of this type can be found in the literature.

To achieve the separation of the groundwater component, a more feasible method would appear to be that based on a direct observation of the groundwater level in a single well, or on a utilisation of the mean groundwater level found by observing several wells. The relationship between the groundwater level and the baseflow is first determined for the low-flow period and then extrapolated (Fig. 3.5). It does



3.5.1 Groundwater level–baseflow relationship based on the observation of deep groundwater circulation.



3.5.2 Groundwater level–baseflow relationship developed as an enveloping curve. The groundwater level is a product of several different régimes.

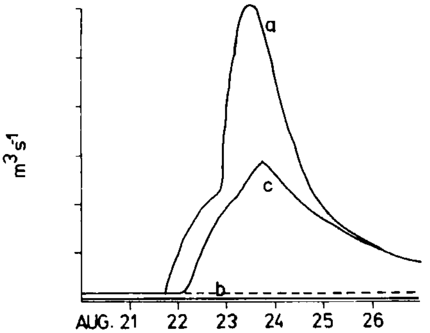
not necessarily follow, however, that for any discharge this relationship will remain linear. For instance, in shallow aquifers the rising limb of the baseflow component may be bent by fast simultaneous outflow (Fig. 3.6).

A method based on the presumption that soils absorb almost all the β-radiation of the percolating water was developed by Balek and Rálková [1]. It has been proved experimentally that the β-radiation of the surface runoff remains equal to the β-radiation of the rainfall, while the β-activity of the subsurface component is absorbed in the soil layers. The subsurface water contribution at any point of the flood can be evaluated from the expression:

$$Q = Q_g + Q_s \tag{3.1}$$

$$Q_a = Q_g a_g + Q_s a_s \tag{3.2}$$

where Q is the flow in the river at a given point, Q_g is the subsurface flow component, a is the specific β-activity of the river water at the same point of the wave, a_g is the specific β-activity of the river water before the river discharge starts to rise, a_s is the specific β-activity of the rainfall; and Q_s is the surface runoff component.



3.6 The rising line of a shallow baseflow component, bent by the impact of fast simultaneous outflow. Thames River, Canada. Deep baseflow – b, shallow baseflow – c, flood wave – a.

Combining the above expressions, we obtain

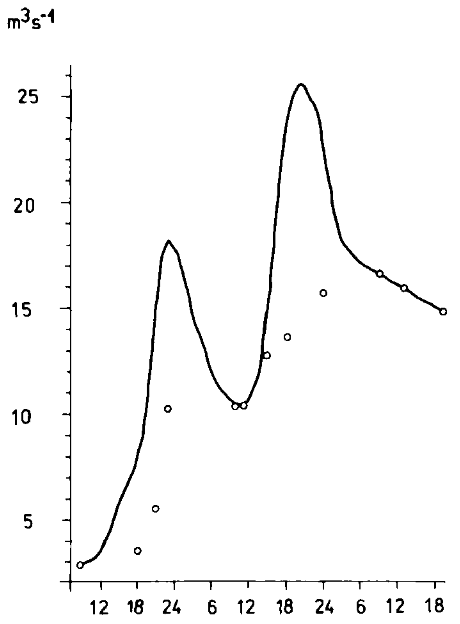
$$Q_s = Q \frac{a - a_g}{a_s - a_g} \tag{3.3}$$

The flood wave of the Volyňka river in southern Czechoslovakia is depicted as an example in Fig. 3.7.

A similar approach was described by Toler [2]. It was based on the analysis of the chemical composition of the rain and subsurface water. Pinder and Jones [3] attempted to apply a similar method to determine the groundwater component of peak discharge.

In 1970 Kille [4] presented a method based on the analysis of the discharge minima. First, he calculated the duration curve from the monthly minima and then the duration curve was replaced by a smoothed curve which appears on log-normal paper as a straight line. Then the annual groundwater outflow is determined on that line as a median (Fig. 3.8).

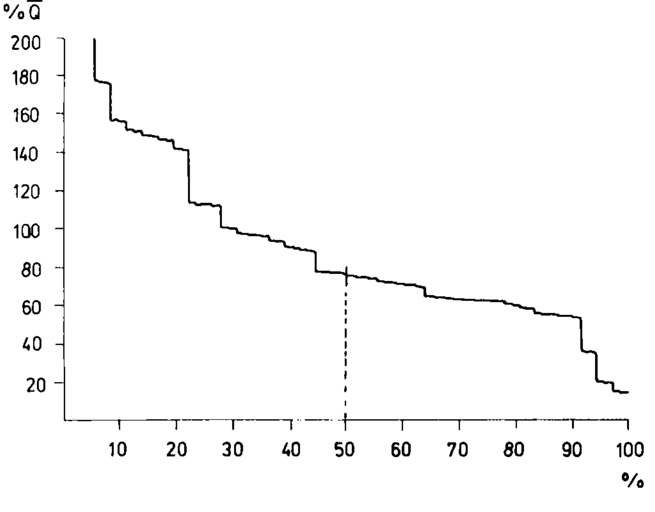
A similar method was introduced by Castany et al. [5] who concluded that the groundwater component could be calculated as an arithmetic mean of the minimum daily discharges on thirty consecutive days.



3.7 Points of separation of the surface runoff as determined by β-activity measurements of the rain and river water. Volyňka River, Czechoslovakia.

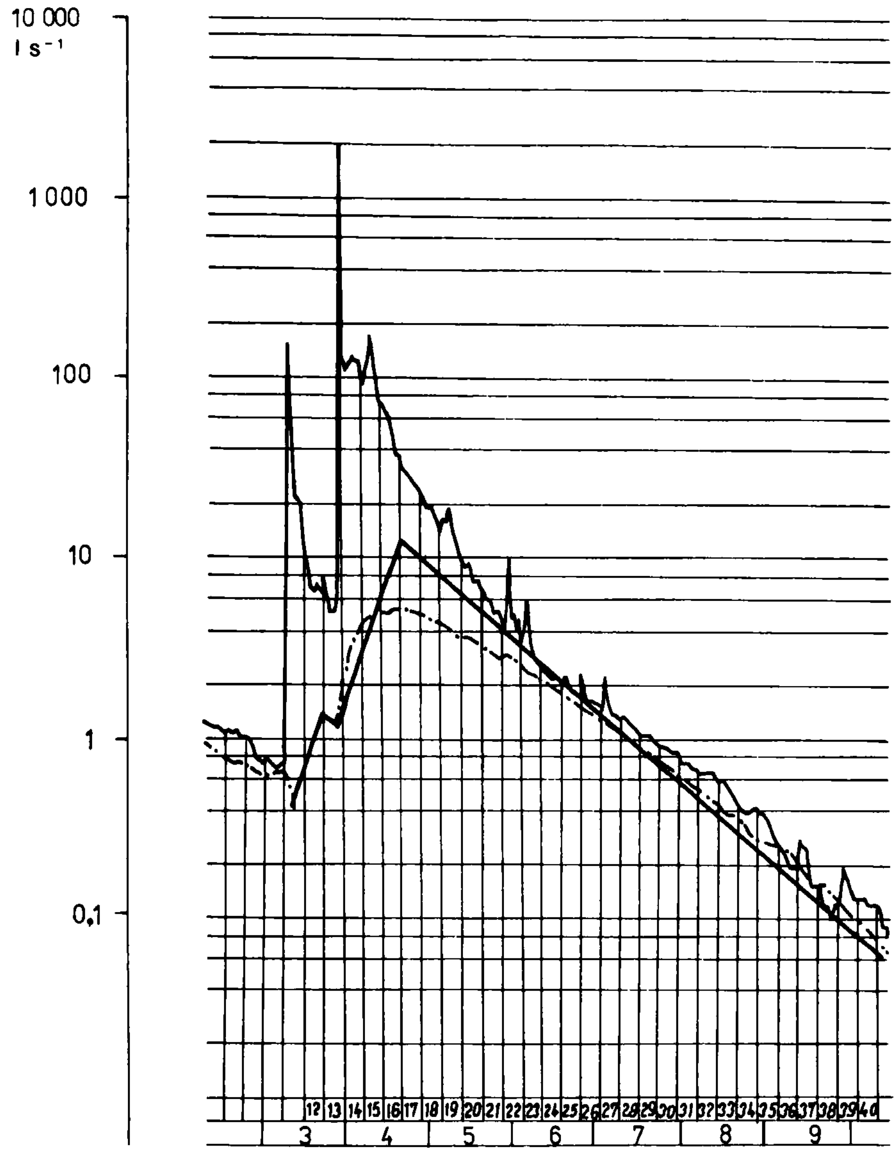
Kříž [6] applied both methods to the minimum flow sequences of Cuban rivers, and in several cases he found great differences between the results given by Kille's and by Castany's method. One reason for this may be the impact of man on the minimum discharge régime and criteria set up artificially.

A simple analytical method for the separation of the groundwater outflow is based on the relationship of those points which on the hydrograph are typical for the baseflow. Those considered typical are the points of annual minima, minima during



3.8 Duration curve of monthly minima and mean annual baseflow according to Kille. Upper Turonian formations, Czechoslovakia.

the growing and non-growing season, and any discharge during prolonged dry periods. In this approach the baseflow component is often overestimated, particularly during the wet years when the contribution from the interflow is rather high throughout the year. It is therefore more convenient to attempt an additional sep-



3.9 An assessment of the recession coefficient on a semilogarithmic scale; $K = 0.9987$ (hourly base) represents shallow outflow. Actual baseflow –·–·–·–, optimized baseflow ——.

aration of the hydrograph representing the dry periods before and after the wet year. Then the final/initial points of the baseflow for the preceding/following period can be utilised.

Another convenient method is based on analysing the shape of the recession curve, defined as

$$Q_t = Q_0\, e^{-\alpha t^n} \qquad 3.4$$

where Q_t is the discharge during the recession period at time t, Q_0 is the discharge at time $t = 0$, e is the base of the natural logarithm, n is the parameter characterising the shape of the recession curve and α is the recession coefficient. A practical form of the recession expression is

$$Q_t = Q_0 K^t \qquad 3.5$$

where K is the recession coefficient.

It should be noted that K depends also on the selected time interval t (hours, days, etc.).

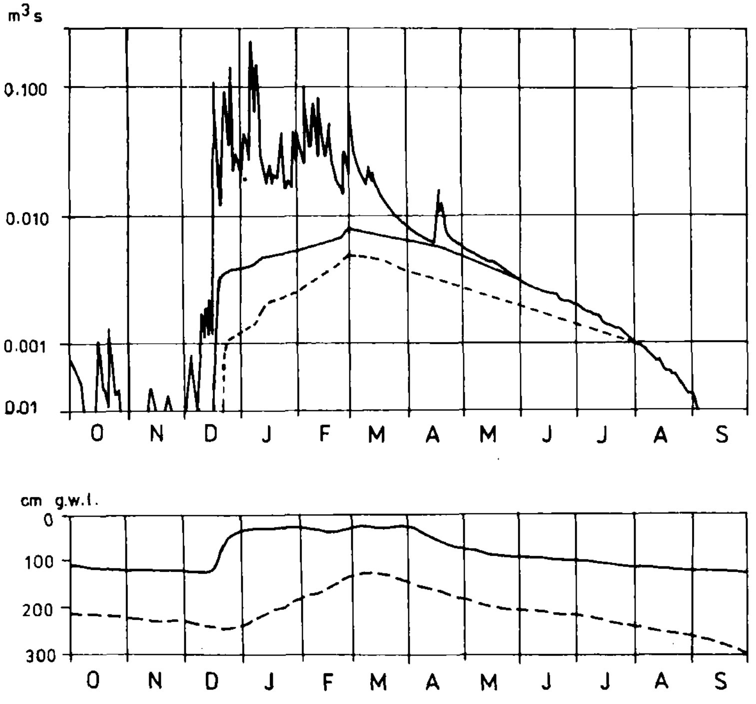
A graphical solution of the above expression is presented in Fig. 3.9.

The term recession curve is sometimes reserved for any descending limb of the hydrograph. Thus components other than the baseflow may be involved. For a recession curve which reflects a steady decrease of the baseflow as a result of the depletion of groundwater resources, the term depletion curve is used. Thus the beginning of the baseflow depletion curve should be allocated to the point where surface runoff and subsurface flow have ceased.

In 1970 Appleby [7] stated that "...the runoff is formed from many sources forming a complex of independent exponential reservoirs set up by an innumerably large series of past event transients, associated with random rainfall episodes". This led to the adaption of the above methods to a three-component separation, which is adequate for many practical purposes. Using hydrograph recession analysis we can often find at least three different components, namely the surface runoff, subsurface flow, and baseflow. An analysis of the hydrograph plotted on log-normal paper is a feasible technique for separating the three components.

Fig. 3.10 provides an example of three-component separation of a hydrograph. On a log-normal scale, the components of the hydrograph which represent an intermittent tropical stream also can be traced using the groundwater level fluctuation in shallow and deep aquifers. A shallow aquifer is fully recharged soon after the beginning of the rainy season and its groundwater level remains almost steady during the rainy season, while in the deep aquifer the groundwater level rises steadily almost to the end of the rainy season. The hydrograph components behave accordingly.

Another method achieving three-component separation is based on the geochemical mass balance and measurements of electric conductivity (Balek et al. [8]). It was found that, under a forest, the effect of the surface runoff is suppressed in favour of the subsurface flow. Here, as the discharge increases the water conductivity also increases because the percolating water is enriched with minerals from the soil



3.10 Three-component separation of the hydrograph in forested/grassland catchment, based on the groundwater level fluctuation under forest - - - - - - -, and under grassland ———. Northern Zambia.

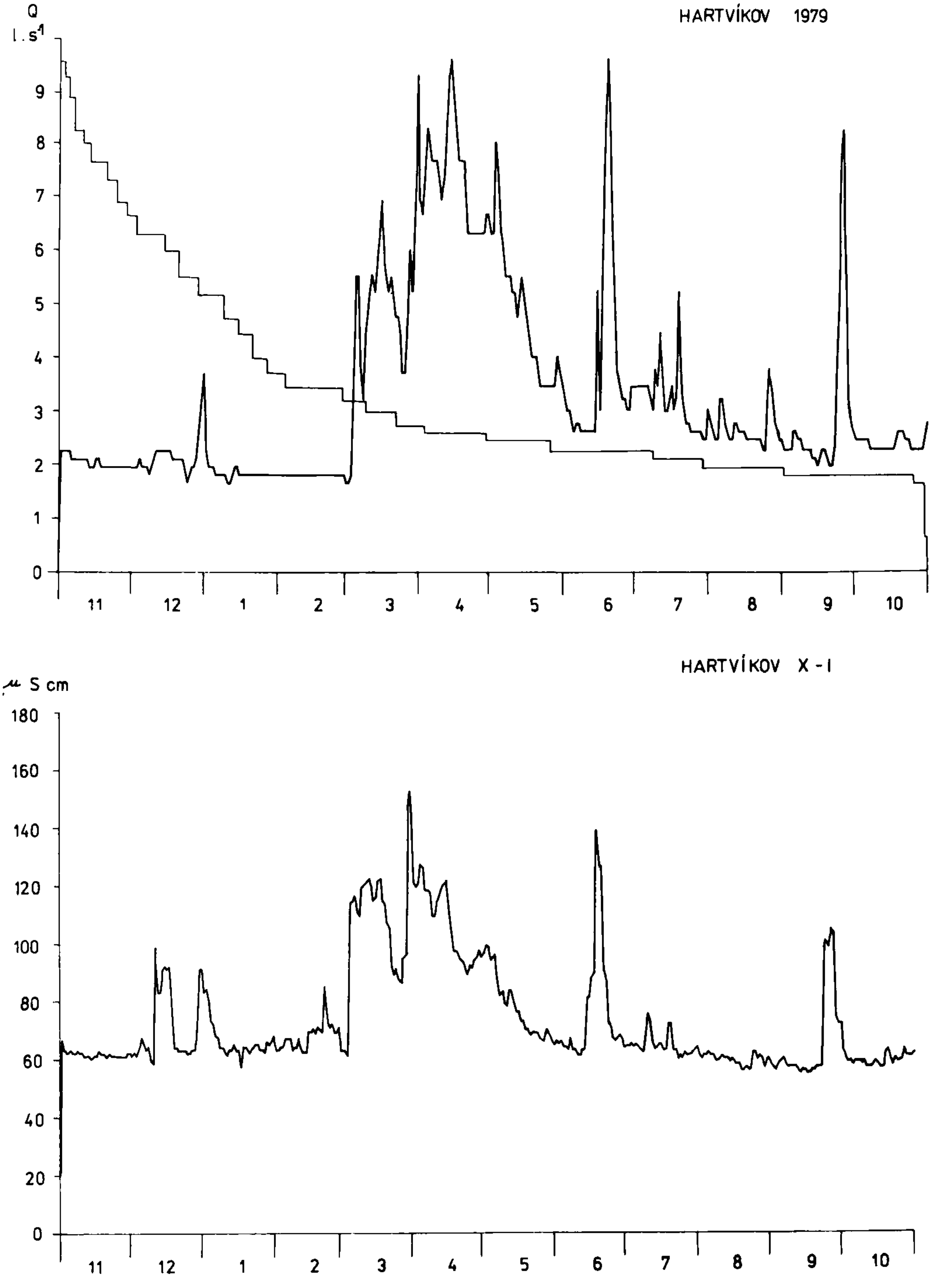
profile (Fig. 3.11). In the agricultural catchment the flood peaks comprising the surface runoff and non-mineralised rain water reduce the conductivity (Fig. 3.12). Separation equations, set up accordingly, have the form:

$$Q = Q_s + Q_g + Q_i \tag{3.6}$$

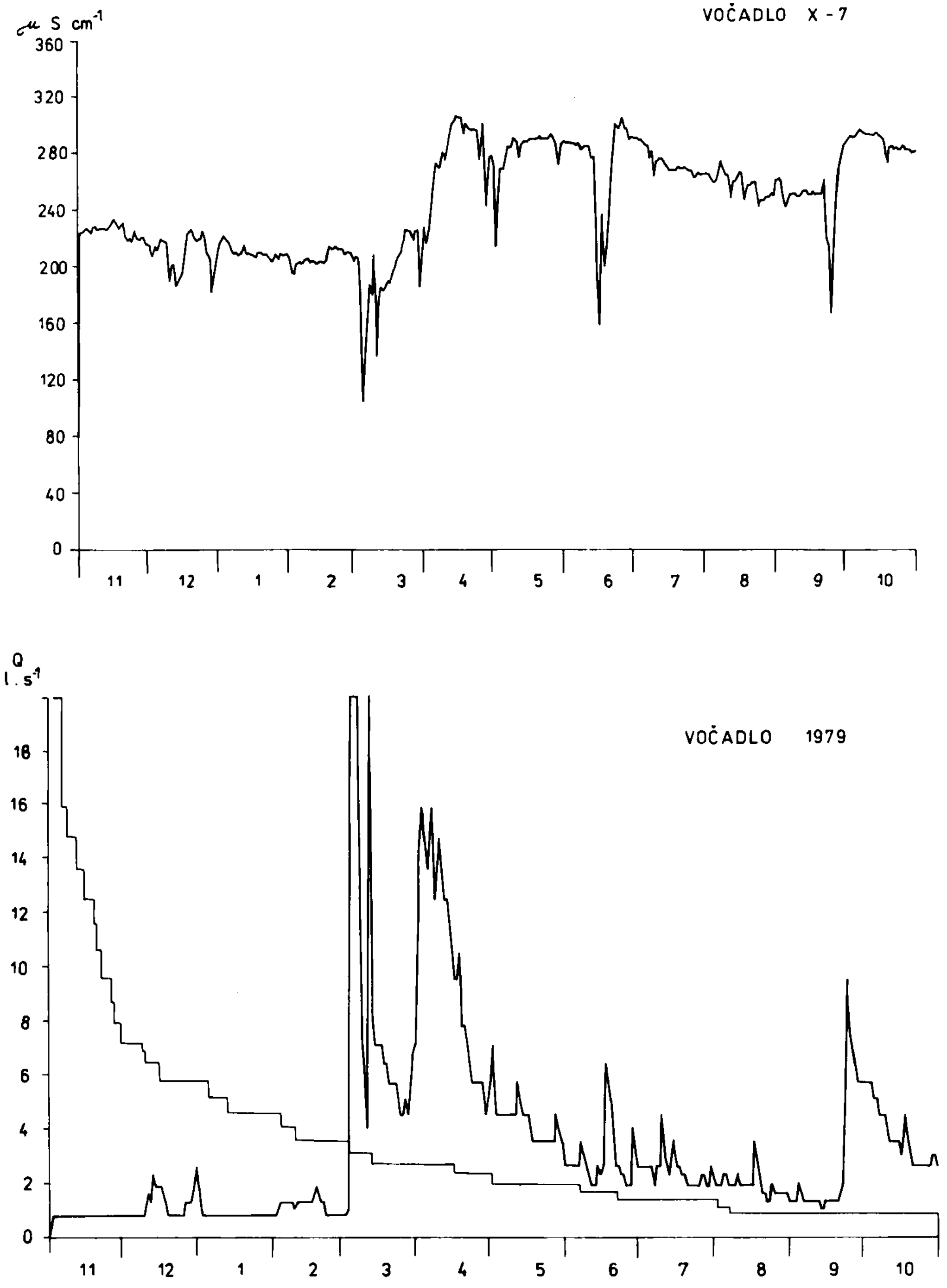
$$Qa = Q_s a_s + Q_g a_g + Q_i a_i \tag{3.7}$$

$$Q_g = 0.28(Hp - S)A \ln K \tag{3.8}$$

Here Q is the instant discharge, in $m^3\,s^{-1}$; Q_s is the surface runoff, in $m^3\,s^{-1}$; Q_g is the baseflow, in $m^3\,s^{-1}$; Q_i is the subsurface component, in $m^3\,s^{-1}$; a, a_s, a_g, a_i are the corresponding conductivities, μS; H is the depth of the groundwater level below the surface, in mm; A is the drainage area of the basin, in km^2; p is the non-capillary porosity of the saturated zone; S is the groundwater zone capacity, in mm; and



3.11 Hydrograph and conductivity fluctuation in a forest catchment, central Czechoslovakia.



3.12 Hydrograph and conductivity pattern in an agricultural catchment, central Czechoslovakia.

K is the recession coeffieicient, based on hourly intervals. Then

$$Q_s = \frac{Q(a - a_i) - [0.28(Hp - S)\,A \ln K(a_g - a_i)]}{a_s - a_i} \tag{3.9}$$

and

$$Q_i = Q - Q_g - Q_s \tag{3.10}$$

As an example, we can cite the calculation of the three components in the Večadlo agricultural catchment in Central Czechoslovakia, for the day when the groundwater level reached the ground surface:

Drainage area = 0.391 km^2,
discharge during the storm = 0.014 81 m^3 s^{-1},
depth of the groundwater table = 0,
noncapillary porosity of the saturated zone = 0.166,
groundwater zone capacity = 154.29 mm,
hourly recession coefficient K = 0.999 81,
conductivity of the river water = 352 μS,
conductivity of the rain water = 33 μS,
conductivity of the interflow measured at the drainage outlet = 767 μS,
conductivity of the baseflow = 308 μS.

Applying the equations we obtain:
$Q_g = 0.001\,9$ m^3 s^{-1},
$Q_s = 0.007\,21$ m^3 s^{-1},
$Q_i = 0.005\,7$ m^3 s^{-1}.

In general, various separation methods can be found to suit the different types of data available; however, without a solid long-term observation of the river discharge and groundwater level, any assessment of the components will supply only approximate results. Another method, based on the so-called duration curves, is discussed under 3.3. and the simulation approach is presented in Chapter 6. The assessment of the baseflow component is closely linked to the determination of the so-called safe yield. This will be discussed in Chapter 4.

Very approximately, the groundwater component calculation is based on the presumption of a steady flow (see subsection 2.8.1):

$$Q = kIMb \qquad \mathrm{m^3\,s^{-1}},$$

where k is the coefficient of permeability, in m s^{-1}; I is the gradient (dimensionless); M is the aquifer's thickness; and b is the length of the cross-section for which the groundwater is calculated, in m.

For the transmissibility coefficient $T = kM$ m^2 s^{-1}; then $Q = TIb$ m^3 s^{-1}.

3.2 SYNTHESIS OF THE WATER BALANCE EQUATION

For a basin defined by hydrological criteria, some of the water balance equations discussed in Chapter 2 can be utilised once the water balance components have been determined. The validity of these equations must be carefully checked, because the accuracy of at least some of the components is rarely certain.

In more complicated structures and/or hydrological régime it may be more difficult to set up a standard form of any water balance equation. Approximately-defined boundary conditions and unusual climatological régimes frequently pose problems. Several examples of water balance equations of this kind can serve as guidance.

3.2.1 Water balance in semi-arid and arid regions

In semi-arid and arid regions, the stream channels carry the flood water only occasionally when erratic rainfall occurs. The water balance equation is then relatively simple. For instance, in a semi-arid basin a mean annual rainfall of 420 mm was observed and the mean annual potential evaporation was 1760 mm. The flood runoff, estimated using channel hydraulics formulae, was 8 mm per year. The water balance equation takes the following form:

Rainfall mm	Surface runoff mm	Loss mm
420	8	412

The equation is based on the presumption that no water is transported out of the basin as subsurface flow beneath dry river beds; otherwise the water balance equation would also contain the groundwater outflow component. Usually, the seasonal fluctuation of groundwater level below the root zone indicates that some active groundwater storage is taking place. Additionally, in some tropical regions the trees are so adapted that they can tap almost all the water which is available in the zone of capillary rise. Many species survive on far less water, which means that some part of the groundwater storage could be utilised. After pumping out, say, 8 mm of water per year in the above basin, the water balance equation becomes:

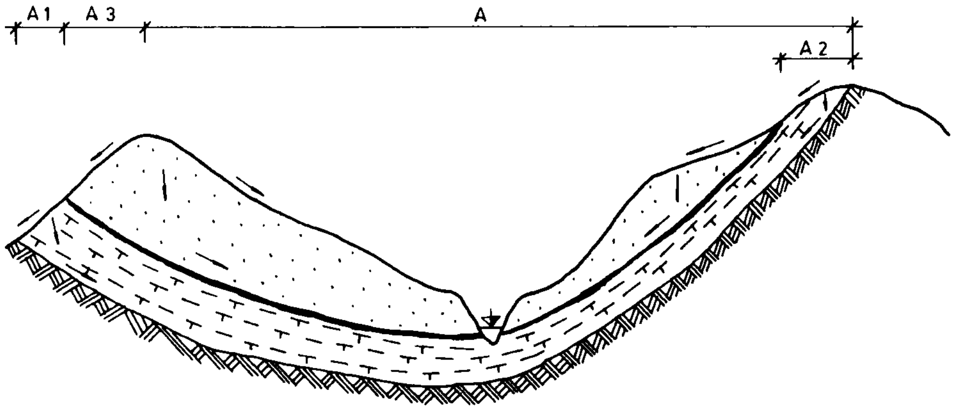
Rainfall mm	Surface runoff mm	Pumping mm	Loss mm
420	8	8	404

We can see that although no groundwater component has been involved in the first equation, exploitation of groundwater is feasible and reduces the amount left for transpiration.

Comparing the two water balance equations, one can expect that changing the conditions in the root zone will cause an environmental effect on the vegetation.

3.2.2 Complicated structures

The water balance can be calculated for basins in which the outcrop of the water-bearing aquifer is relatively small in comparison with the whole basin. Sometimes the outcrop is located outside the hydrological boundaries. A standard water balance equation may supply misleading results, particularly when it is used for a comparison with the water balance of a neighbouring basin. As an example the hydrological basin in Fig. 3.13 is presented. The alluvial formations form an extensive plain which is drained by the main stream. To the hydrological drainage area A,



3.13 A complicated hydrogeological structure with a different hydrological and hydrogeological divide.

comprising alluvial deposits, a subarea A_3 also contributes. Draining into the main stream there is also a groundwater system with a deep circulation, recharged at the outcrop A_2 (within the hydrological boundaries) and at outcrop A_1 outside them. The total observed outflow is $0.921\ m^3\,s^{-1}$, the annual precipitation amounts to 580 mm, the drainage area A is $200\ km^2$, $A_1 = 25\ km^2$, $A_2 = 30\ km^2$ and $A_3 = 10\ km^2$. Thus the hydrogeological drainage area is $A + A_1 + A_3 = 235\ km^2$.

Separating the hydrograph components of the annual hydrograph we obtain:

Surface runoff	$0.090\ m^3\,s^{-1}$,
subsurface flow	$0.220\ m^3\,s^{-1}$,
baseflow	$0.630\ m^3\,s^{-1}$.

The corresponding values for the hydrological boundaries are:

Surface runoff	14.19 mm,
subsurface flow	34.69 mm,
baseflow	99.34 mm.

The standard water balance equation can be written as:

Precipitation mm	Surface runoff mm	Subsurface flow mm	Baseflow mm	Loss mm
580	14.19	34.69	99.34	431.78

This water balance equation tells very little about the recharge process. Therefore, for hydrogeological purposes the equation can be written separately for the deep and shallow regions.

Because the baseflow is recharged only from the area $A_1 + A_2 = 55\,km^2$, we can write for the deep circulation:

Precipitation mm	Surface runoff mm	Deep baseflow mm	Loss mm
580	14.19	361.23	204.58

For the shallow circulation, which occurs from the area $A - A_2 + A_3 = 180\,km^2$,

Precipitation mm	Surface runoff mm	Shallow outflow mm	Loss mm
580	14.19	38.54	527.27

Comparing the two equations, we can conclude that in the first case the soil layers of the outcrop are highly permeable and the aquifer, having many fissures, permits a fast transport of the infiltrating water under the root zone. Thus an immediate return of the infiltrating water to the atmosphere is more hindered than in the case of the alluvial water, where the area is more exposed to the process of evapotranspiration.

3.2.3 Communication impact

Very peculiar water balance results can be obtained in karstic regions, where the communication component may play a significant rôle. An example of this is the water balance of the Bystrica river in southern Albania. Here a mean annual discharge of $21.77\,m^3\,s^{-1}$ was observed from the hydrological drainage area of size $46.32\,km^2$. The mean annual rainfall is 1758 mm and thus the water balance equation can be written as:

Precipitation mm	Runoff mm	Loss mm
1758	14 822	−13 064

Here it is obvious that the "loss" component has nothing in common with the evapotranspirational loss. To find the actual impact of the communication component, another gauging station was established downstream, where from a drainage area of $106.25\ km^2$ a discharge of $23.90\ m^3\ s^{-1}$ was observed. Thus a mean annual discharge of $2.13\ m^3\ s^{-1}$ was determined from the area of $59.93\ km^2$ and for the precipitation of 1808 mm.

Separating the surface runoff component in the synthetised hydrograph, obtained as the difference between the downstream and upstream hydrographs, the water balance equation for the area delimited by the two gauging posts will be:

Precipitation mm	Surface runoff mm	Groundwater outflow mm	Loss mm
1808	184	937	687

We can expect that the local groundwater contribution in the upper basin is approximately $1.38\ m^3\ s^{-1}$ and the long-distance contribution is $20.39\ m^3\ s^{-1}$. Thus the size of the recharge area outside the basin limits can be calculated:

$$A = \frac{20.39 \cdot 3600 \cdot 24 \cdot 365}{1000 \cdot 937} \doteq 686.3\ km^2.$$

This information can help, for instance, when the recharge area has to be defined in order to delimit the groundwater protection zones.

Provided we intend to take into account different rainfall patterns, the rainfall-runoff relationship has to be developed for the region and the runoff component modified in accordance with changing precipitation.

3.2.4 Complex water balance equations

In some experimental areas several water balance components are measured with a high accuracy and some additional measurements are available. Here a comparison of basic and complex water balance equations can serve as a guidance when the groundwater régime is analysed. As an example let us use water balance data from an experimental base in northern Czechoslovakia. In the year 1983 the mean annual precipitation was 630.34 mm, surface runoff from the grassland 2.70 mm, from the bare soil 9.36 mm and groundwater recharge in the lysimeter was 37.25 mm. Transpiration from grass was determined as 121.37 mm, from oak 120.77 mm and from spruce 95.88 mm. Potential evaporation was calculated with the Penman formula using the mean daily values of meteorological phenomena, and estimated at 510.35 mm. Discharge observed at the outlet was $0.044\,8\ m^3\ s^{-1}$, the soil moisture increment was 7.2 mm, and the groundwater deficit -5.6 mm.

The depth of the water-bearing aquifer is about 50 metres below the stream bed, transmissibility is $4.23\ m^2\ s^{-1}$ and the dip of the impervious bed approximately

1‰. The hydrological and hydrogeological drainage areas are 5.88 and 7.245 km^2, respectively. Groundwater dating in deep strata indicates the time of residence as 6000 years; combined C^{14} and tritium analysis of the stream water indicates that about 30% of the water is of recent origin, while 70% of the water comes from ancient sources.

Total annual runoff from the hydrogeological drainage area is 195 mm, and for the annual surface runoff of 6.03 mm we obtain the groundwater component as 188.97 mm. If 30% of this water is of recent origin, this means that 56.69 mm has been recharged recently and 132.28 mm a long time ago. Because the deep aquifer within the catchment limits is well below the stream bed, we can assume that the old water has been recharged outside the hydrogeological catchment through long-distance communication.

The subsurface outflow under the stream bed was calculated as 9.2 mm, and this water, too, is quite possibly old. Thus the water balance for 1983 can be written (in mm):

$$\overset{P}{630.34} + \overset{C_0^i}{141.48} = \overset{O_0^b}{132.28} + \overset{O_r^b}{56.69} + \overset{C_0^s}{9.2} + \overset{\Delta S}{7.2} - \overset{\Delta G}{5.6} + \overset{R}{6.03} + \overset{E_t}{566.02}$$

where P represents precipitation, C_0^i inflow communication (132.28 + 9.2), O_0^b baseflow of old water, O_r^b baseflow of recent water, C_0^s outflow under the river bed; ΔS is the soil moisture increment; ΔG the groundwater deficit, and R the surface runoff. Total evapotranspiration exceeds the potential evaporation, and consists of the transpiration and evaporation from the surface of the vegetation and soil. The mean transpiration was estimated as 112.67 mm, and thus the evaporation is 569.22 − 112.67 = 456.55 mm. The annual recharge measured on a single lysimeter corresponds reasonably well with the groundwater outflow of recent water.

A simple water balance equation for the same area has the form (in mm):

$$\overset{P}{630.34} = \overset{O}{195} + \overset{E_t}{435.34}$$

which obviously is far less informative from the point of view of the formation of groundwater resources; for some purposes it might even supply misleading results.

3.3 DURATION CURVE ANALYSIS

In the hydrological approach, the mean annual discharge or its equivalent (in millimetres) provides fundamental information on the watershed behaviour. However, from the water balance equation we obtain no information on the distribution of the runoff process. In order to assess the discharge distribution, the percentage of time in which each observed discharge is equalled, or exceeded, is determined from what is called the duration curve. In fact, this curve is an indicator of the basin potential for a given period. To find out on how many occassions the variate has a higher or lower value than a certain given value, one uses the so-called cumulative frequency, obtained as a summation over all time intervals. For this purpose, the daily discharges are arranged in order of their magnitude. The amplitude of this array is defined as the difference between maximum and minimum values. This amplitude is divided into preselected intervals. The number of occurrences in each of them is called frequency "*n*". When divided by the total number of variates, *n* in the series becomes the relative frequency "*f*". It is graphically represented by a frequency distribution

Tab. 3.1 Mean daily discharges of an intermittent tributary of the Kafue river, Zambia, 1950 – 1960

	Oct.	Nov.	Dec.	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sep.
1	0.08	0.03	0.03	0.22	0.14	0.22	0.22	0.08	0.08	0.08	0.08	0.03
2	0.08	0.03	0.03	0.22	0.14	0.22	0.22	0.08	0.08	0.08	0.08	0.03
3	0.08	0.03	0.03	0.20	0.14	0.22	0.22	0.08	0.08	0.08	0.08	0.03
4	0.08	0.03	0.03	0.14	0.14	0.22	0.22	0.08	0.08	0.08	0.08	0.03
5	0.08	0.03	0.03	0.20	0.14	0.22	0.22	0.08	0.08	0.08	0.08	0.03
6	0.08			0.25	0.14	0.22	0.22	0.08	0.08	0.08	0.08	0.03
7	0.08			0.25	0.14	0.22	0.22	0.08	0.08	0.08	0.08	0.03
8	0.08			0.22	0.14	0.22	0.22	0.08	0.08	0.08	0.08	0.03
9	0.08		0.08	0.20	0.22	0.22	0.22	0.08	0.08	0.08	0.08	0.03
10	0.08		0.08	0.14	0.22	0.22	0.20	0.08	0.08	0.08	0.08	0.03
11	0.08		0.08	0.14	0.22	0.22	0.20	0.08	0.08	0.08	0.08	0.03
12	0.08		0.08	0.14	0.22	0.22	0.20	0.08	0.08	0.08	0.08	0.03
13	0.08	0.03	0.08	0.14	0.28	0.22	0.20	0.08	0.08	0.08	0.08	0.03
14	0.03	0.08	0.14	0.14	0.34	0.22	0.14	0.08	0.08	0.08	0.08	0.03
15	0.03	0.08	0.31	0.14	0.28	0.22	0.08	0.08	0.08	0.08	0.08	0.03
16	0.03	0.08	1.09	0.14	0.25	0.22	0.08	0.08	0.08	0.08	0.08	0.03
17	0.03	0.08	1.99	0.15	0.22	0.22	0.08	0.08	0.08	0.08	0.08	0.03
18	0.03	0.08	0.20	0.15	0.22	0.22	0.08	0.08	0.08	0.08	0.08	0.03
19	0.03	0.08	0.17	0.15	0.22	0.25	0.08	0.08	0.08	0.08	0.08	0.03
20	0.03	0.08	0.62	0.15	0.22	0.25	0.08	0.08	0.08	0.08	0.08	0.03
21	0.03	0.11	2.07	0.15	0.22	0.25	0.08	0.08	0.08	0.08	0.08	0.03
22	0.03	0.11	0.22	0.15	0.22	0.25	0.08	0.08	0.08	0.08	0.08	0.03
23	0.03	0.11	0.22	0.15	0.22	0.25	0.08	0.08	0.08	0.08	0.08	0.03
24	0.03	0.11	0.25	0.15	0.22	0.25	0.08	0.08	0.08	0.08	0.08	
25	0.03	0.11	0.48	0.25	0.22	0.25	0.08	0.08	0.08	0.08	0.08	
26	0.03	0.08	0.28	0.17	0.22	0.25	0.08	0.08	0.08	0.08	0.08	
27	0.03	0.08	0.22	0.17	0.22	0.25	0.08	0.08	0.08	0.08	0.08	
28		0.03	0.22	0.17	0.22	0.25	0.08	0.08	0.08	0.08	0.06	
29		0.03	0.22	0.17	0.22	0.25	0.08	0.08	0.08	0.08	0.03	
30		0.03	0.22	0.17		0.25	0.08	0.08	0.08	0.08	0.03	
31			0.17	0.14		0.22		0.08		0.08	0.03	

bar hydrograph (histogram) from which the duration curve can also be obtained graphically. An example of the calculation of the duration curve from data in Tab. 3.1 is in Tab. 3.2. A graphical interpretation appears in Fig. 3.14.

As can be seen, the lower part of the duration curve provides information on the fair weather flow or, in other words, on the contribution of the groundwater discharge. Indeed, sometimes the extrapolated lower part of the duration curve can serve as a guide for the assessment of the groundwater flow (Fig. 3.14).

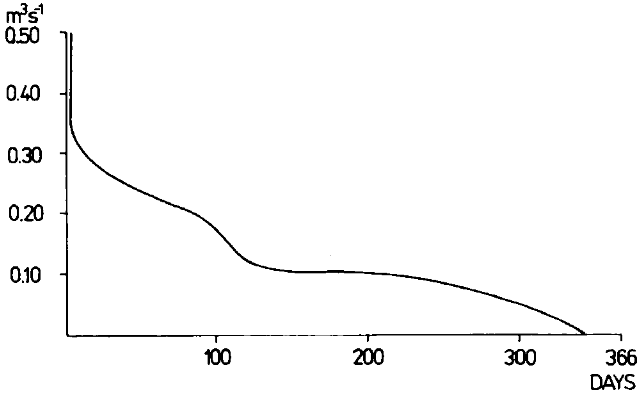
A comparison of the shapes of several duration curves provides a basis for the comparison of groundwater régimes of neighbouring basins. For such a purpose the curves are sometimes plotted on a dimensionless scale, so that instead of the discharge values alone the ratio discharge/mean annual discharge is used.

In Fig. 3.15 duration curves are plotted which represent typical river/groundwater régimes in the Tertiary-Cretaceous formations. A steep slope of curve No. 1 indicates that during the low flow discharge period the groundwater outflow either was not stabilised or was rather low. On the other hand, a substantial part of curve No. 2 is nearly parallel with the axis x and thus the régime, with the exception of a few months, is dominated by the groundwater outflow. The curves Nos. 3 and 4 represent river régimes with a moderate groundwater contribution. The gradient of the curves informs us that the groundwater component is close to a discharge which is likely to exceed it on 300 days per year. A tangent to the bottom part of both curves can be consulted, when the baseflow component is to be determined.

The subregional duration curves can also be used to determine the drainage areas pertinent to the water balance equations which have been set up for each subregion. Such a problem may occur if the hydrological régime of the region is a product of several different groundwater régimes such as deep and fast shallow circulations, or when the preferred pathways have to be allocated.

It can be written

$$\sum_{I=1}^{n} q_I^{p} A_I^{p} + \sum_{J=1}^{m} q_J^{s} A_J^{s} + \sum_{K=1}^{o} q_K^{g} A_K^{g} = Q \qquad 3.11$$



3.14 Duration curve representing an intermittent stream (data given in Tab. 3.1 and 3.2). The baseflow contribution is well below 0.1 m^3 s^{-1}. Middle Zambezi basin.

Tab. 3.2 Calculation of the duration curve

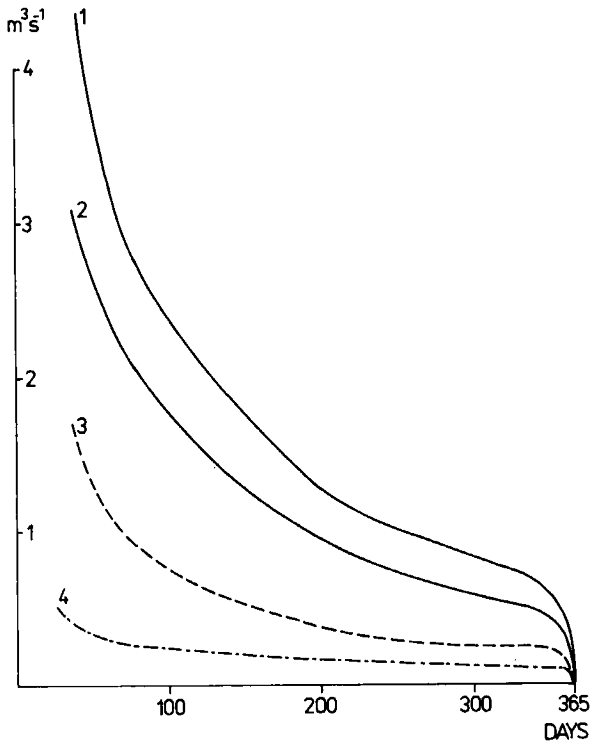
River: Nakambala | Duration of daily discharge year 1959 – 1960 | Station: bridge

Discharge interval $m^3 s^{-1}$	Month	Number of days when discharge was equal to or greater than in the first column and equal to or less than that shown in the second column													Cumulated	
		10	11	12	1	2	3	4	5	6	7	8	9	Totals	Totals	% of time
>.50				3										3	3	0.8
0.31 – 0.50				2		1								3	6	1.6
0.21 – 0.30				9	6	20	31	9						75	81	22.1
0.15 – 0.20				3	16			4						23	104	28.4
0.11 – 0.14			5	1	9	8		1						24	128	35.0
0.06 – 0.10		13	9	5				16	31	30	31	28		163	291	79.5
0.01 – 0.05		14	9	5								3	23	54	345	94.2
0.00		4	7	3									7	21	366	100.0
		31	30	31	31	29	31	30	31	30	31	31	30	366		

where n is the number of the subregions with different surface runoff régimes, m is the number of subregions with shallow groundwater flow pattern, and o is the number of subregions with different deep circulations; q_I^p is the surface runoff yield ($m^3 s^{-1} km^{-2}$) from a subregion A_I^p (km^2); q_J^s is the subsurface outflow yield ($m^3 s^{-1} km^{-2}$) from a subregion A_J^s (km^2); and q_K^g is the baseflow yield ($m^3 s^{-1} km^2$) from a subregion A_K^g (km^2). A is the drainage area of the region. Indexed areas are called partial, active or contributing areas. For regions with autochthonous sources:

$$A = \sum_{I=1}^{n} A_I^p \qquad 3.12$$

Q is the total runoff from the region ($m^3 s^{-1}$).



3.15 Duration curves representing the régimes of four neighbouring rivers in Tertiary-Cretaceous formations, Czechoslovakia.

For the drainage areas A_J^s it is expected that

$$\sum_{J=1}^{m} A_J^s = XA \qquad 3.13$$

where X is the percentage of subregions from which a subsurface outflow can be expected, such as alluvial deposits, terraces, Quaternary formations, slopes, and fissured rocks with shallow soil profile. Then it follows that

$$\sum_{K=1}^{o} A_K^g = A(1 - X) \qquad 3.14$$

For regions with allochthonous sources:

$$\sum_{J=1}^{m} A_J^s + \sum_{K=1}^{o} A_K^g > A \qquad 3.15$$

The drainage areas A_I^p are determined by the hydrological division of the subregions. Q is based on direct observation and/or on convenient measurements in the ungauged outflow/inflow areas. The q_J and q_K values are obtained from the water balance equations so that for each contributing area

$$q(\mathrm{m^3\,s^{-1}\,km^{-2}}) = \frac{1000\,O\ (\mathrm{mm})}{3600 \cdot 365 \cdot 24}$$

Coefficient X can be estimated from hydrogeological topographical and soil maps. For the determination of A_K^g we can set up expressions as, for instance:

$$\begin{aligned} &\frac{q_1^g A_1^g}{q_2^g A_2^g} = k_{1,2}, \ldots, \frac{q_1^g A_1^g}{q_K^g A_K^g} = k_{1,K} \\ &\vdots \\ &\frac{q_K^g A_K^g}{q_1^g A_1^g} = k_{K,1}, \ldots, \frac{q_K^g A_K^g}{q_{(K-1)}^g A_{(K-1)}^g} = k_{K,(K-1)} \end{aligned} \qquad 3.16$$

The guidance for the determination of the coefficents k is the comparison of the lower parts of the duration curves, typical for the subregions. Usually the coeficients are estimated as a ratio of Q values likely to be in excess for 270–300 days. The number of k values should be sufficient for the calculation of A_K^g values, and in some cases also for the calculation of A_J^s values.

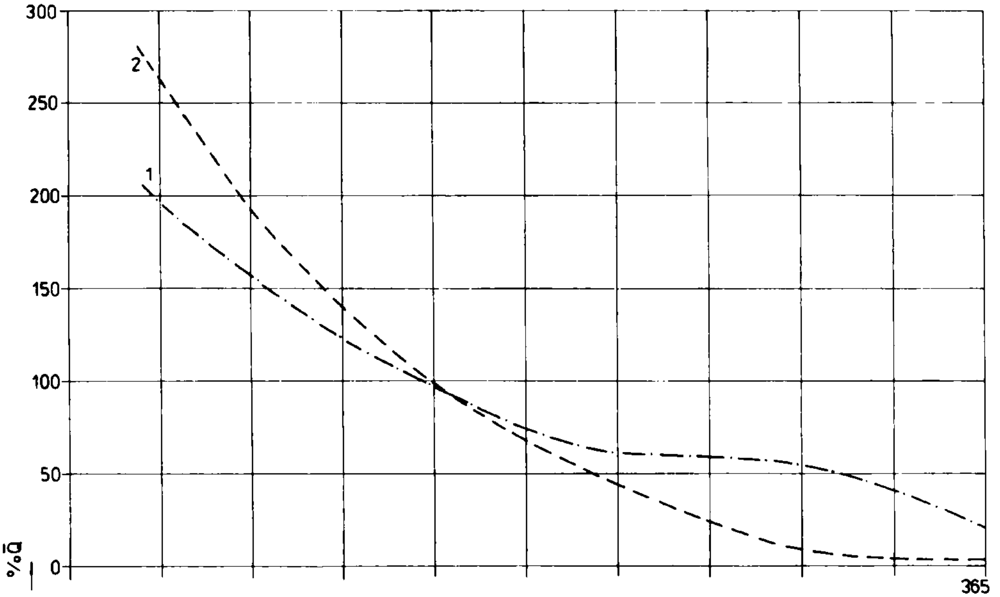
Characteristic shapes of some duration curves reflect also the geological features of the subregions. In Fig. 3.16 is an example of two duration curves typical for two basins on the Balkan Peninsula. Curve No. 1 is typical for a basin formed by limestone and dolomitic formations with a highly permeable soil layer, while the curve No. 2 is characteristic for basins with prevailing volcanic rocks with a less permeable surface (Balek [9]).

It should be noted that geological features may dominate the duration curves of extensive basins, but for small catchments the shape of the duration curves can be influenced by other features. A stabilised duration curve, for instance, can betray a contribution from some communication component or the impact of a lake régime.

Mimikou and Kaemaki [10] synthetised flow duration curves of Greek rivers by using third-order polynomials of the type

$$Q = a - bD + cD^2 - dD^3 \qquad 3.17$$

where Q is the discharge, D is the corresponding time of excess and a, b, c, d are constants. These curves may be found suitable, perhaps except for minimum discharges.



3.16 Characteristic duration curves for two Balkan basins with different geological features; Albania.

3.4 STABILISATION OF THE WATER BALANCE COMPONENTS

The hydrological cycle, and particularly the groundwater outflow components, may become temporarily unbalanced under artificial recharge and pumping. A depletion of the groundwater storage produces a decrease in the baseflow. An additional effect, that of a temporarily decreasing recharge in dry years, may become another destabilising factor, particularly when the baseflow becomes very slight.

In order to estimate the behaviour of the destabilised system, first a simplified water balance is written:

$$P = R + O + E_t \qquad 3.18$$

where P is the long-term mean annual precipitation, R the surface runoff, O the groundwater outflow and E_t the evapotranspiration (all values in mm year^{-1}).

In fact O also indicates the long-term mean annual recharge, provided there has been no artificial recharge or depletion. After a depletion or recharge of M mm year^{-1} occurs, a new value Z mm year^{-1} is introduced:

$Z = O - M$ for depletion,
$Z = O + M$ for recharge.

For such an event the water balance equation has to be rewritten. The groundwater outflow component O_N (mm year^{-1}) occurring N hours after the beginning of the event will be:

$$O_N = K^{(N-1)}[8760G_i(1 - K) - Z] + Z \qquad \text{mm year}^{-1} \qquad 3.19$$

Here G_i is what is called the initial groundwater storage of the active part of the groundwater resources (in mm) and K is the recession coefficient based on hourly intervals. For G_i it can be written:

$$G_i = \frac{3.6Q_{gi}}{(-\ln K)A} \qquad 3.20$$

where Q_{gi} is the estimated initial baseflow (in m^3 s^{-1}), K is the recession coefficient and A is the drainage area (in km^2).

It should be noted that this baseflow is rather an instant value corresponding to the N-th hour, but, in the pertinent water balance equation, written for the N-th hour, all values are in mm year^{-1}.

As well as an instant baseflow, we may need to estimate the mean groundwater outflow $\bar{O}_T$ (in mm year^{-1}) during a period of T hours:

$$\bar{O}_T = \frac{1}{T}\left\{[8700G_i(1 - K) - Z]\frac{1 - K^T}{1 - K}\right\} + Z \qquad \text{mm year}^{-1} \qquad 3.21$$

Finally, the time T_Q, after which the baseflow is equal to a certain value O_T mm year^{-1}, can be estimated:

$$T_Q = \left\{\frac{\log\left[\dfrac{Q_T - Z}{8760G_i(1 - K - Z}\right]}{\log K} + 1\right\}\frac{1}{8760} \text{ years} \qquad 3.22$$

These approximate expressions are valid for conditions of steady recharge and/or depletion and, therefore, deviation from the predicted values can be foreseen particularly in extremely dry or wet periods. As can be seen from the expressions, G_i should be determined for the undisturbed initial conditions, when $O \doteq Z$.

As an example let us examine the stabilisation of the régime in the Lower Kamenice basin, drained by a tributary of the Elbe at the Czech-German borders. The water balance of the basin, based on observations made between 1931 and 1960, is per year:

Precipitation mm	Surface runoff mm	Baseflow mm	Evapotranspiration mm
800.4	21.82	357.08	421.50

The drainage area is 40.87 km^2, an initial baseflow in 1931 was 0.473 m^3 s^{-1} $\sim$ $\sim$ 364.78 mm year^{-1}. The recession coefficient K based on the hydrograph analysis and for hourly intervals is 0.999 83. Thus initial groundwater storage $G_i = 244.95$ mm.

The planned pumping rate for the basin is 0.252 m^3 s^{-1}, an equivalent of 194.45 mm year^{-1}. Thus

$$Z = 357.08 - 194.45 = 162.63 \text{ mm}$$

The reduced mean annual baseflow for the same period 1931 – 1960 will be:

$$O_T = \frac{1}{30 \cdot 8760}\left\{[244.95 \cdot 8760 \cdot 0.000\,17 - 162.63]\frac{1}{0.000\,17}\right\} + $$
$$+ 162.63 = 167.15 \qquad \text{mm year}^{-1}$$

The stabilisation of the system is expected after T years when the baseflow O_T will be in balance with Z. We can solve the problem approximately under condition that the difference between the baseflow O_T and water extraction Z will be negligible, or

$$O_t - Z \doteq \frac{Z}{8760} = 0.018$$

Then

$$T = \left\{\frac{\log\left[\dfrac{0.018}{244.95 \cdot 8760 \cdot 0.000\,17 - 162.63}\right]}{\log 0.999\,83} + 1\right\}\frac{1}{8760} = 6.26 \text{ years}$$

The mean water balance equation for the period 1931 – 1960 and for the pumping 0.252 m^3 s^{-1} is:

Precipitation mm	Surface runoff mm	Baseflow mm	Pumping mm	Groundwater deficit mm	Evapo-transpiration mm
800.4	21.82	167.15	194.45	−4.52	421.5

The water balance for the stabilised régime after 6.26 years is:

Precipitation mm	Surface runoff mm	Baseflow mm	Pumping mm	Evapotranspiration mm
800.4	21.82	162.63	194.45	421.5

It was to be expected that the mean groundwater deficit in the water balance for 1931 – 1960 would be higher at the beginning of the period and would diminish after seven years, provided that from the hydrological point of view each individual year was normal.

3.5 THE CONCEPT OF PROBABILITY IN GROUNDWATER RÉGIME ANALYSIS

Principles of probability analysis, which have been widely applied in the analysis of surface hydrology phenomena, can also be applied to the solving of groundwater problems. For instance, an empirical probability curve can be used when the probability of occurrence of a certain groundwater level, considered as critical, is to be determined. The following expressions are usually applied:

$$p = \frac{m - 0.5}{n} 100\%$$

$$p = \frac{m}{n + 1} 100\% \qquad 3.23$$

$$p = \frac{m - 0.3}{n + 0.4} 100\%$$

where p is the probability of occurrence, %; m is the number of events in the descending or ascending sequence, and n is the number of events.

Usually the period of observation of groundwater levels is rather short, and in order to assess the levels corresponding to low probabilities of occurrence some theoretical probability curve has to be applied. A wide assortment of theoretical probability curves can be found in the specialised literature. As in surface hydrology, Pearson's type III distribution curve was found suitable for the analysis of groundwater phenomena.

Tab. 3.3 contains empirical and theoretical probabilities of occurrence of the lowest mean annual groundwater level in the borehole Banín in northern Czechoslovakia. It is based on observations made during the period 1931 – 1960 (Fig. 3.17).

By arranging the same sequence in ascending order we obtain another piece

Tab. 3.3 Deepest/highest groundwater levels below surface likely to be exceeded in the Banín borehole, northern Czechoslovakia

Probability that the deepest level is likely to be exceeded, %	1.0	5.0	10.0	50.0	90.0	95.0	99.0
Probability that the highest level is likely to be exceeded, %	99.0	95.0	90.0	50.0	10.0	5.0	1.0
Empirical, m		24.20	21.10	17.10	16.70		
Theoretical, m	24.43	23.03	22.21	19.76	17.66	17.0	16.26

of information – the highest mean annual groundwater level likely to be reached.

The probability concept need not be used only for the analysis of groundwater levels. For instance, such an analysis of the discharge minima which are observed

3.17 Groundwater level fluctuation in the Banín borehole, 1932–1960.

in rather extensive basins will provide general information on the probability of occurrence of active groundwater resources.

Tab. 3.4 shows the probability of occurrence of the minimum discharges of the River Oubangui in the Central African Republic. Knowing the drainage area to be approximately $A = 500\,000\ \mathrm{km}^2$ in size, and taking the recession coefficient K, shown by the hydrograph analysis, as 0.9996 we can estimate the lowest active groundwater storage likely to be reached once in 100 years:

$$G = -\frac{Q_{100}^{\min} \cdot 3.6}{A \ln K} = 7.0\ \mathrm{mm}\ \text{ or }\ 3.5\ \mathrm{km}^3.$$

Tab. 3.4 The minimum discharge $\mathrm{m}^3\,\mathrm{s}^{-1}$ which is likely to be exceeded in the River Oubangui, Central African Republic

Probability:	1.0	5.0	10.0	20.0	50.0	90.0	95.0	99.0
Minimum Q:	390	405	560	650	820	1 030	1 200	1 500

Similarly, for 99 years out of 100, the active groundwater storage will be 27.0 mm or 13.5 km^3.

Extended examples of the application of the probability concept are found in Chapters 7 and 8.

3.6 REFERENCES

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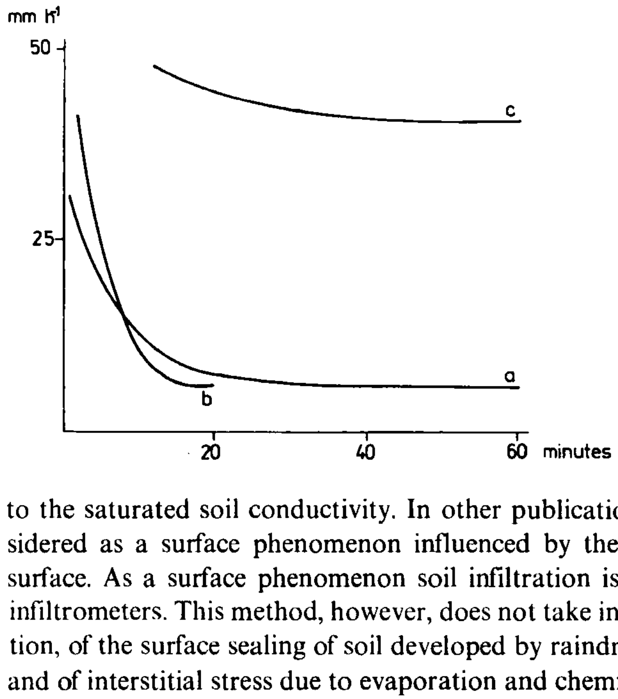
4 GROUNDWATER RESOURCES FORMATION AND THE SAFE YIELD CONCEPT

4.1 INFILTRATION, PERCOLATION AND SEEPAGE

When water is ponded on the soil surface, the rate of entry of water into the soil is initially very high, but drops after a short period of time to a steady state called the final infiltrability. Infiltration curves characterising this process are given in Fig. 4.1.

In some publications we find that the infiltration process is defined as the result of the permeability of the entire soil profile, and the final infiltrability is closely related



4.1 Soil infiltration curves for loam soil. Bare soil – *a*, continuous corn – *b*, grass/legume – *c*.

to the saturated soil conductivity. In other publications the process has been considered as a surface phenomenon influenced by the noncapillary porosity of the surface. As a surface phenomenon soil infiltration is often measured by means of infiltrometers. This method, however, does not take into account the effect of vegetation, of the surface sealing of soil developed by raindrops or dispersed soil particles, and of interstitial stress due to evaporation and chemical crusting.

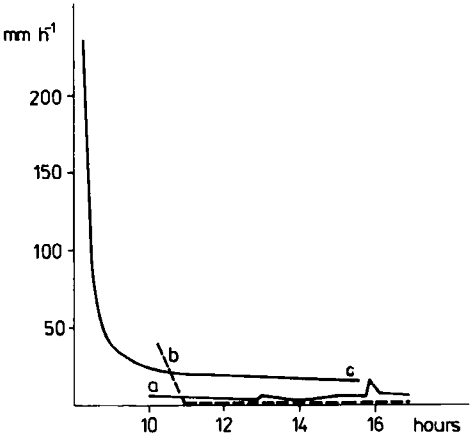
The crust is often made up of two parts: a very thin non-porous layer and a layer up to 5 mm thick of inwashed fine particles. Thus the crust may become a decisive factor in the infiltration process.

Obviously the infiltration process is influenced by the soil texture and structure. Bare soil, for instance, may soak up the first portion of heavy rain and then will swell and become waterproof. Water infiltration into the same soil which is covered by thick vegetation is delayed by the process of interception, and the soil can absorb more water; thus seasonal changes of soil infiltration can occur in agricultural fields.

Soil moisture plays an equally important rôle. The rate of infiltration is high when soil is dry, but later on, after the capillary pores have been filled, it drops. One of the basic infiltration formulae was developed by Philip [1].

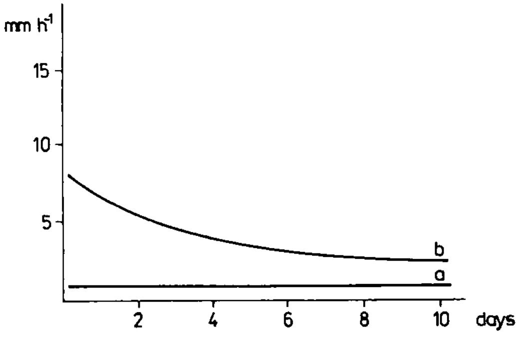
$$f(t) = 0.5St^{-0.5} + A \qquad \mathrm{cm\,s^{-1}}$$

where A is a value close to the saturated soil conductivity, in cm s^{-1}; S is sorptivity, in cm $s^{-0.5}$ and $f(t)$ is the current rate of infiltration, in cm s^{-1}, at time t.



4.2 Infiltration into virgin soil. Summer season – a, frozen soil – b, thawing soil – c. After Zavodchikov.

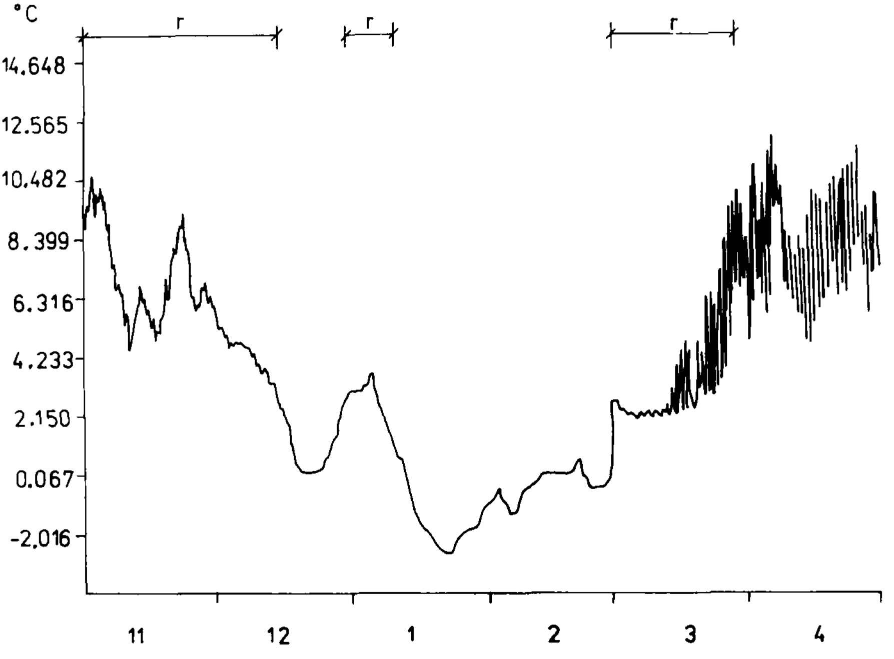
The infiltration conditions change significantly when soil is frozen. The water content of frozen soil plays a particularly important rôle. This should be taken into account in temperate regions where a considerable portion of groundwater recharge originates from thawing snow. Where the snowpack has accumulated on wet frozen soil, a great part of potential groundwater recharge flows away as surface runoff. Fig. 4.2 shows Zavodchikov's graph [2] of infiltration in unfrozen soil, frozen soil and thawing soil. Gorbunov [3] presented graphical comparison of the infiltration into saturated and unsaturated soil as thawing begins (Fig. 4.3). Fig. 4.4



4.3 Infiltration into alluvial soil as thawing begins. Saturated soil – a, unsaturated soil – b. After Gorbunov.

shows the pattern of soil temperature fluctuation around 0 °C and groundwater recharge at the experimental base Nedamov, Czechoslovakia.

For groundwater studies it is recommended to derive the infiltration characteristics from a comparison of actual and net rainfall, by analysing rainfall and runoff records. The result should be considered as a catchment infiltration, which, because it incorporates an interception effect, is not comparable with the soil infiltration.



4.4 Groundwater recharge during the period of soil thawing; Nedamov experimental base, Czechoslovakia. Soil temperature measured 10 cm below surface. Recharge period — *r*.

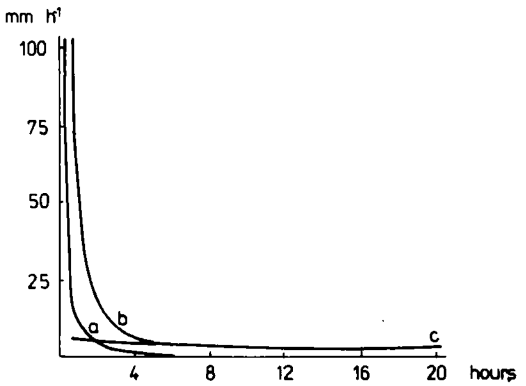
The terms used for the infiltration phenomena in soil physics and in catchment hydrology often differ. Therefore it can be an advantage to consider the results based on rainfall-runoff analysis, as representative for the whole catchment. As an example, the result of such an analysis from the mountainous area of Krkonoše, Czechoslovakia [4], is presented in Fig. 4.5. The process was studied there under conditions of high previous rainfall, and was compared with catchment infiltration in Texas.

The following equation for catchment infiltration has been developed [27]:

$$f(t) = (f_m + f_b) \exp\left[- \frac{\log(f_m + f_b) - \log 2f_b}{T \log e} \right] t - f_b \qquad 4.1$$

where $f(t)$ is current catchment infiltration rate at time t, f_m is maximum infiltration rate into the soil, f_b is the percolation rate within the soil profile and t is the period of time after which the current catchment infiltration rate is close to final infiltrability.

While infiltration is a surface phenomenon, the quantity of water which passes a given point in the soil profile in a given time is sometimes called percolation. This process occurs when the soil moisture content exceeds field capacity. In some works the process is also called the transmission rate, while other authors prefer to work with the term saturated soil conductivity. Todd [5] gave the term „percolating rate" a legal background, because in the past some water laws recognised as percolating waters "those moving slowly through the ground but not a part of any definite underground stream", contrary to "water sources" either on the surface or underground.



4.5 Catchment infiltration process; Texas woodland – *a*, Texas cropland – *b*, forest/pasture mixture, high antecedent rainfall, Czechoslovakia – *c*.

Seepage is a term used for the movement of water through a porous media to the ground surface or into surface water bodies, or vice versa. Storm seepage is that part of the rainfall which infiltrates into the soil surface and moves laterally through the soil horizons directly toward the streams as ephemeral, shallow or perched groundwater above the main groundwater level. The process is sometimes called subsurface flow.

4.2 NATURAL RECHARGE OF GROUNDWATER

Natural recharge of groundwater may occur by precipitation, or from rivers, canals and lakes. Practically all the water from these sources is of meteoric origin. What is known as juvenile water (of volcanic, magmatic and cosmic origin) only exceptionally contributes to the groundwater recharge.

The process of recharge is very complicated. Perhaps one of the most important factors is the time delay between the time when the meteoric water enters the soil profile, and the time when it is manifested as an effectively exploitable groundwater source.

In principle, the following types of natural groundwater recharge can be recognised (Balek [33]):

1. Short-term recharge, which may occur occasionally after a heavy rainfall, mainly in regions without marked wet and dry seasons.

2. Seasonal recharge, which usually occurs regularly, e.g. at the beginning of the snowmelt period in temperate regions, or during the wet period in wet and dry regions. Occasionally, when there has been an unfavourable development of soil moisture, the recharge may be poor or may not occur at all.

3. Perennial recharge, which may occur in humid regions with an almost permanent downward flow of water.

4. Historical recharge, which occurred a long time ago and contributed to the formation of the present groundwater resource. This phenomenon is closely linked with what is known as groundwater residence time. The residence time is defined as the time which has elapsed between the time at which a given volume of water was recharged and the time when it reaches the groundwater table (Campana and Simpson [6]). Sometimes the time taken by a given volume of water to be transformed into baseflow is decisive for the assessment of the residence time.

A simple assessment of the annual recharge is given in a Unesco textbook [7]:

$$R = SAh \qquad 4.2$$

where R is the volume of recharge, A is the groundwater fluctuation area, S is the storage coefficient, and h is the mean amplitude of the fluctuation. It is stated that a lack of precision in determining S may significantly reduce the accuracy of the assessment.

The effect of the vertical recharge of historically and/or regionally exogeneous water may also significantly influence the estimate. In practice this problem was analysed in the Kalahari desert by De Vries [8]. He concluded that an active recharge can take place only in areas where the thickness of the sand cover is not less than 6 metres, otherwise seasonal soil moisture retention and subsequent evapotranspiration would consume the penetrating water. Under such circumstances the current depletion is taken from groundwater storage formed during a major wet period 4000 years ago. The relationship between the historical recharge and the residence time was analysed by Campana and Simpson [6], with the help of the discrete state compartment model, used in conjunction with environmental isotope distributions as calibration data.

Sophocleous and Perry [9] measured the recharge processes in Kansas, U.S.A., on unconsolidated deposits without any surface drainage pattern. They concluded that previous moisture conditions, thickness of the aquifer, and the nature of the unsaturated zone, were major factors affecting recharge. They observed a significant variation in recharge, ranging from less than 2.5 mm to approximately 154 mm in late winter and spring.

Sharma and Hughes [10] studied the groundwater recharge in Western Australia by investigating environmental chloride; they concluded that about 50% of recharge occurs through preferred pathways which bypass the soil profile. However, being unable to quantify the considerable variability in the recharge and preferred pathways, they propose further research into this problem.

Many hydraulic models are based on the presumption of a constant recharge into the aquifer. However, as some experiments have shown (Kitching et al. [11]), groundwater recharge may be limited to several weeks or days in a year, even under humid climatic conditions, and to very short periods in arid regions. As stated by Kitching et al.: "... input data to groundwater models consist of the hydrogeological framework of the aquifer including boundary conditions, abstractions, outflows and recharge; however, in many cases the estimation of recharge constitutes one of the largest errors introduced into the model". Besbes and de Marsily [12] attempted to define and quantify the difference between infiltration and recharge in an aquifer. They concluded that average infiltration and average recharge are identical over a long time period, and the distinction accounts only for the delay and smoothing that its percolation through the unsaturated zone imposes in transforming infiltration into recharge. It should be added, however, that there is a great difference between the actual infiltration and the potential infiltration which is indicated by infiltrometers. By introducing the theory of unit hydrograph in groundwater hydrology, Besbes and de Marsily found the following relationship between recharge q and water table recovery s:

$$s_j(t) = \int_0^{\infty} q(t - \tau)\, \Psi_j(\tau)\, d\tau \qquad 4.3$$

where $s_j(t)$ is effective recovery of the water table at point j due to the recharge q; $q(t)$ is recharge to the aquifer as a function of time; and $\Psi_j(\tau)$ is unit response of the aquifer to recharge at point j. The effective recovery is defined as the difference between the observed piezometric head in the aquifer, and the hypothetical head d which would have been observed if the recharge had been zero at the time when the calculation started.

Obviously a problem related to the process of groundwater recharge concerns the type of aquifer or system of aquifers involved. In general, it is assumed that in unconfined aquifers the prevailing type of recharge is a rather rapid process in which water of recent meteoric origin reaches the piezometric head in the whole region at once. On the other hand, in confined aquifers it is assumed that the meteoric water has entered the recharge area far away from the location where it will be exploited, or that recharge occurred during the period when the hydraulic properties of upper formations, which at present serve as an aquiclude, were more favourable.

Intensive studies of the recharge process began shortly after isotope methods were introduced to groundwater hydrology. Whereas, in the hydraulic approach, a direct recharge from meteoric water is assumed (perhaps with some more or less constant inflow as a boundary condition), by using radioactive methods it has been proved frequently that the recharge system is often much more complicated.

In order to measure direct recharge from rainfall, the most effective method seems to be the water balance calculation based on lysimetric experiments. Because the lysimetric measurements are representative for single points within the region, their validity for more extensive areas has to be carefully examined.

Kitching et al. [11] took the lysimetric approach in the Bunter sandstone formations, Nottinghamshire, and found that 135 mm of the 600 mm rainfall became groundwater recharge. Morel and Wright [13] published a 10-year record of groundwater recharge for the Chalk in West Suffolk, with results which coincide with the above (Tab. 4.1).

In Bunter sandstone, however, Bath, Edmunds and Andrews [14] analysed the water of ancient origin and concluded that a large proportion of water in the Bunter formations was recharged already during the late Pleistocene and early Holocene, in a period when rainfall was delivered from continental air masses. They concluded that water quality resulting from the mixing of non-polluted water and recently-polluted water provides useful information on the paleoclimatology.

Tab. 4.1 Typical values of monthly recharge into aquifer in eastern England, after Morel and Wright

Year	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sep.	Oct.	Nov.	Dec.	Total
1965	0	0	12	10	0	0	0	0	0	0	24	91	137
1966	20	54	0	9	0	0	0	0	0	0	24	67	174
1967	23	36	0	12	21	0	0	0	0	0	0	40	132
1968	39	22	1	0	0	0	0	13	56	24	28	42	225
1969	57	49	37	0	30	0	0	0	0	0	0	0	163
1970	49	50	23	29	0	0	0	0	0	0	0	0	151
1971	70	10	23	0	0	0	0	0	0	0	2	15	120
1972	51	30	24	0	0	0	0	0	0	0	0	0	105
1973	0	0	0	0	0	0	0	0	0	0	0	0	0
1974	0	46	2	0	0	0	0	0	0	0	88	21	157
1975	54	13	64	22	0	0	0	0	0	0	0	0	153
1976	0	0	0	0	0	0	0	0	0	0	0	17	17

In Czech sandstone formations, north of Prague, it was found by lysimetric experiments that of 600 – 800 mm of rainfall less than 10% contributed to a groundwater recharge through loess soil. By comparing measurements of tritium and C^{14} it was found that only 30% of the baseflow was of recent origin, while about 70% was up to 6000 years old. The recharge process of recent water can be seen in Tab. 4.2, the soil moisture conditions favourable for the recharge in Fig. 4.6.

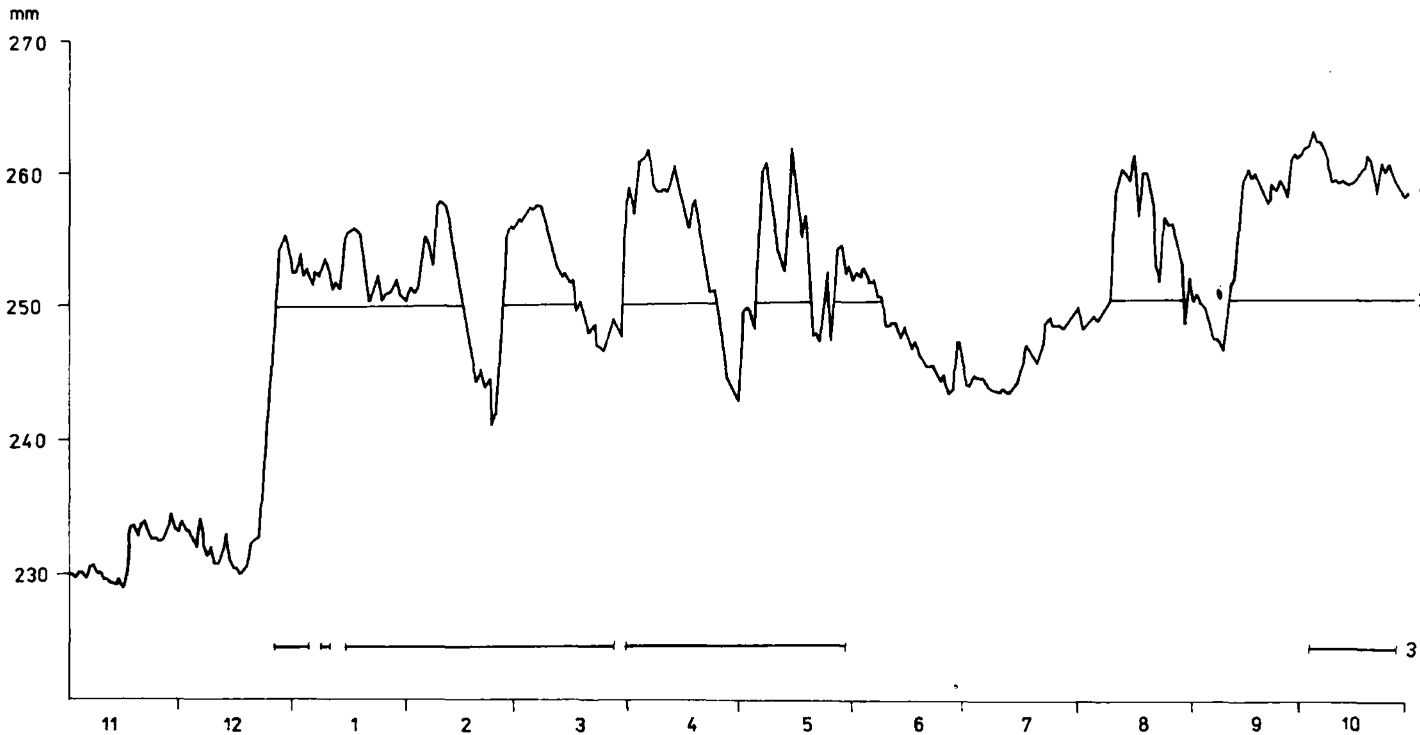
Tab. 4.2 Recharge to Turonian aquifer through loess soil profile in northern Czechoslovakia

Hydrol. year	Date	Duration days	Recharge mm
1982	1. 11. – 15. 12.	45	20.28
	2. 1. – 10. 1.	9	9.12
	2. 3. – 29. 3.	28	11.28
	7. 8. – 11. 8.	5	3.48
			44.16
1983	5. 1. – 14. 4.	100	35.74
	2. 5. – 17. 5.	16	1.51
			37.25
1984	23. 12. – 4. 1.	13	0.57
	8. 1. – 11. 1.	4	0.03
	14. 1. – 27. 3.	58	20.99
	31. 3. – 28. 5.	59	18.45
	3. 10. – 28. 10.	26	3.61
			43.65
1985	24. 11. – 12. 12.	19	4.52
	31. 12. – 4. 1.	6	0.13
	1. 2. – 15. 2.	15	10.89
	27. 2. – 29. 4.	62	12.63
	4. 5. – 22. 5.	19	1.12
	17. 7.	1	2.42
	17. 8. – 1. 9.	16	1.43
			33.14

These results also indicate that in some cases the groundwater level fluctuation itself is not entirely suitable for studying all sources of groundwater recharge. Convenient additional information can be obtained from a graph of soil moisture content. Fleetwood and Larsson [15] analysed the soil moisture fluctuation graphs, together with the rainfall records and groundwater level fluctuation for an area in Southern Sweden. From such a family of graphs a periods of replenishment and depletion of the groundwater resources can be identified. When the amount of soil moisture is not always constant during the period of rising groundwater level, some other sources of groundwater recharge can be expected.

According to the soil moisture fluctuation graph as given in Fig. 4.6 and similar graphs for other years, recharge in Czechoslovakia occurs mainly during the winter period, with some minor periods of recharge in a rainy summer. The total duration of recharge periods is about 80 – 120 days per year. The values given by Morel and Wright [13] indicate that recharge in the U.K. may continue for 4 – 8 months, although during exceptionally dry periods it may cease for a period of 18 months.

Even in arid regions some recharge can be expected. Dincer et al. [16] proved in central Saudi Arabia that about 20 mm of rainfall had annually infiltrated the



4.6 Soil moisture fluctuation in sandy loam, Czechoslovakia – *1*; periods of potential recharge process – *2*; occurrence of actual recharge – *3*. The occurrence of actual recharge during the periods of lower moisture intervals is accounted for by the drainage of the lower layers of the soil profile.

dunes to a depth of at least 7 metres; however, there was no indication whether the recharge had reached the deeper aquifers.

Only with the application of isotope techniques has it become possible to study the recharge process on the wide regional and historical scales. Perhaps the largest residence time of water, equal to 350 000 years, was found by Airey et al. [17] in the Great Artesian Basin of Australia. These authors concluded that a steady piston flow approximation adequately described the groundwater transport of meteoric water, and that for a confined aquifer the contribution from leakage was negligible. By measuring changing chloride levels they were able to prove that the minimum chloride level coincides with the last glacial period, and maxima can be correlated with the relatively warmer periods in the Holocene and during the last Interglacial.

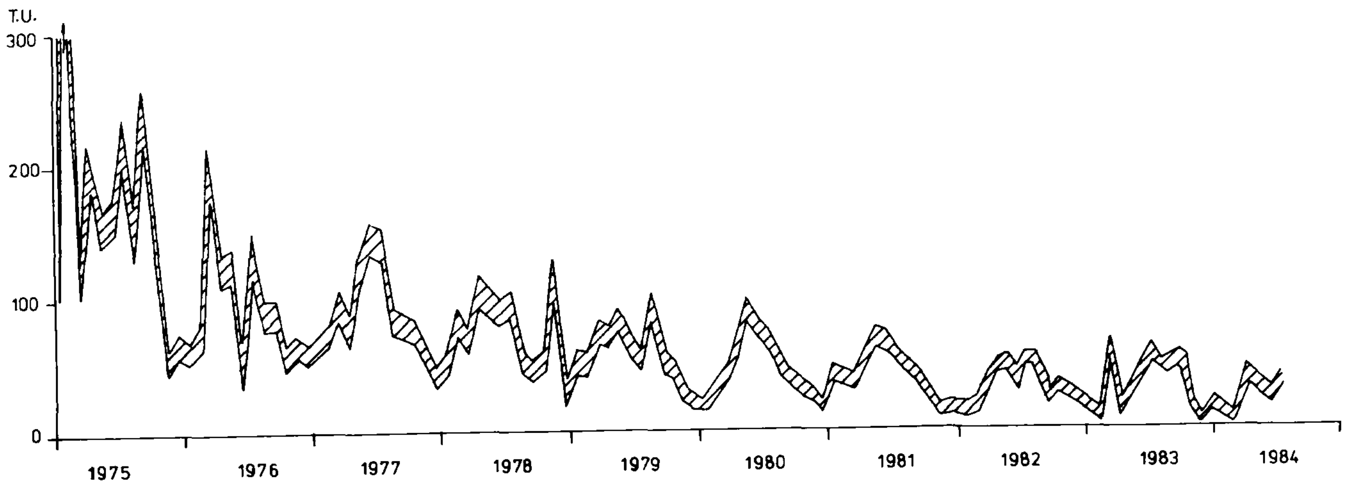
In Sahara it is commonly agreed that groundwater recharges occurred earlier than 20 000 BP, and then between 10 000 – 2 000 BP. The recharge concepts for the period between 20 000 and 10 000 BP are contradictory, and interpretations range from an arid to a humid climate for that period.

Geyh and Backhaus [18] concluded for Central Europe that, during the glacial period, the earth's surface was sealed by permafrost and groundwater recharge was interrupted. Thus, the groundwater recharge here occurred during the Weichselian interstadials up to 20 000 BP, and after 12 000 – 10 000 BP the recharge was large enough to keep the hydrostatic pressure of confined aquifers high.

When water samples taken from the baseflow give a C^{14} analysis indicating water of old age and at the same time contain some tritium, it can be concluded that the groundwater live storage interacting with the baseflow is a mixture of recent and old water. From the size of the storage area between the stream and the outcrop of the water-bearing aquifer and from the baseflow, it is possible to calculate whether the recharge process of the old water is still active, or whether only the remnants of the old water contribute to the baseflow. Such an approach has significance for the long-term strategy of groundwater utilisation.

This method of groundwater dating can be traced back to 1952 when Libby [21] presented the first method of dating archaeological samples. Münnich [19] based the determination of water age on Libby's method. In 1970 Vogel [20] found that only 80 – 90% of C^{14} in bicarbonates of northwestern Europe was of organic origin and that the Münnich's results had to be modified by correcting factors of regional validity. Other methods of correction have been published since then, but, from the hydrological point of view, the separate identification of old and recent water is usually satisfactory.

While C^{14} remains a powerful indicator of the old water's age, tritium has become a very convenient tracer for determining the age of recent water. The mean tritium content in water, given in tritium units T.U., till 1952 was 6 ± 1.5 T.U. After the nuclear explosions in the atmosphere the amount measured in samples reached 10^4 T.U. However, since then it has been steadily decreasing (Fig. 4.7).



4.7 Tritium content in precipitation in 1975 – 1984. Kopisty, Czechoslovakia.

After water has been stored under the ground surface, its tritium content does not increase, but is steadily reduced by decay. Knowing, that the halflife of tritium is 12.26 years, the method can be effectively applied to determine the age of water which is less than 50 years old. In fact, samples which had infiltrated before the period of nuclear explosions in the 1950s have a tritium concentration of less than 2 T.U.

If laboratory tests indicate an old age for a baseflow water sample, and at the same time a significant amount of tritium is found in the sample, certain corrections are needed to obtain correct values. For that purpose it is necessary to determine in the baseflow the age of water which does not contain any tritium. Then

$$C_1 P_t + C_2(100 - P_t) = 100A \tag{4.4}$$

where C_1 is the percentage of C^{14} in the sample without tritium, C_2 is the percentage of C^{14} in recent water, P_t is the volumetric percentage of the old water in the sample and A is the percentage of C^{14} which has been measured in the sample. The volumetric percentage of recent water P is

$$P = 100 - P_t \tag{4.5}$$

Using these results together with the tritium analysis we can estimate the actual value of $\overline{\text{T.U.}}$ in the recent water:

$$\overline{\text{T.U.}} = \frac{100\text{T.U.}}{P} \tag{4.6}$$

where $\overline{\text{T.U.}}$ is the actual amount of tritium and T.U. is the amount found in the sample.

For instance, let us suppose that in the Turonian aquifer the age of the old water was found to be 6921 $\pm$ 390 BP, which corresponds to 42.2% of C^{14}. In the baseflow sample originating from that Turonian aquifer there was determined 72.6% of "modern" C^{14} and 21 $\pm$ 7 T.U. The percentage of C^{14} in recent water was taken as 150–170%.

$$42.2P_t + 160(100 - P_t) = 100 \cdot 72.6$$

and

$$P_t = 74.21\%$$

Then the actual amount of tritium units in the remaining 25.79% is

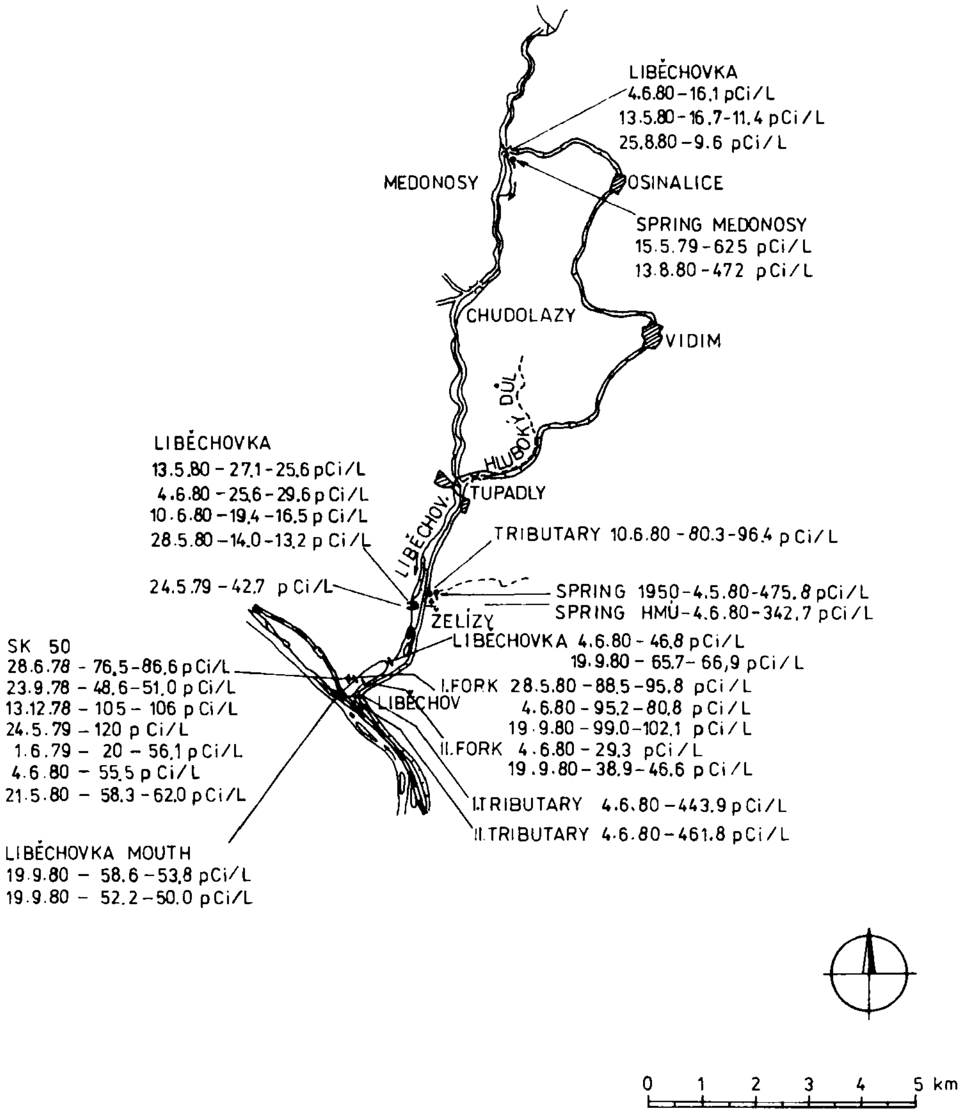
$$\frac{100 \cdot (21 \pm 7)}{25.79} = 54 \sim 109\overline{\text{T.U.}}$$

Let us assume that in the catchment we find the following water balance equation:

$$630P = 180O + 25R + 425E_t$$

where P stands for rainfall, in mm; O for groundwater outflow, in mm; R for surface runoff, in mm; and E_t for loss, in mm. Taking the ratio of old and recent water as calculated, we can write:

$$630P + 133I_g = 47O_r + 133O_g + 25R + 425E_t$$



4.8 Rn^{222} content in samples taken in the Liběchovka basin, Czechoslovakia. A higher Rn^{222} content in the downstream samples indicates a contribution by groundwater having a different origin than water in the upper reaches.

where I_g is the inflow of old water into the catchment, O_g is the outflow of old water, and O_r is the outflow of recent water.

Providing a simplified water balance equation can be applied for a neighbouring catchment, where a contribution of old water cannot be expected, we obtain:

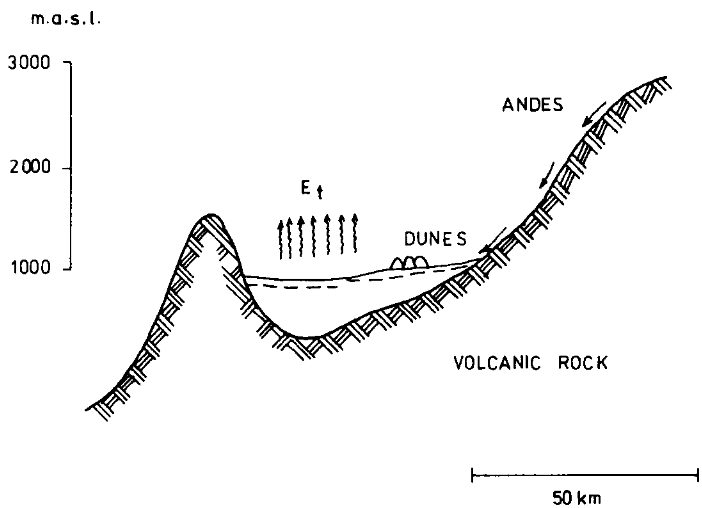
$$630P = 47O_r + 25R + 558E_t$$

This example serves to indicate how carefully the regional validity of water balance equations must be examined, and how the groundwater dating method can be used for such a purpose.

Greater complications occur when the sample contains a mixture of water from two different aquifers. Here the results of chemical analysis can provide additional information. As an example, the measurement of the chloride level in the Australian Great Artesian Basin has been given above. In some cases also the natural radionuclides such as Ra^{226}, U^{234} and Rn^{222} can be utilised. Magri and Tazzioli concluded that in an aquifer with a uniform distribution of radium the water with a higher flow rate may have a higher content of Rn^{222}. Mazor [22] used Ra^{226} and Rn^{222} to interpret the hydrogeological régime of the Great Rift in Africa.

As an example of the application, the measurement of Rn^{222} in Liběchovka basin, Czechoslovakia, is given. Fig. 4.8 gives the results of the measurement of Rn^{222} in the main stream and in the tributaries and outlets of springs in fissured areas. The points of higher Rn^{222} content coincide remarkably with the fissures in the aquiclude between Turonian and Cenomanian aquifers. This approach facilitates the assessment of another recharge component in the water balance calculation.

In arid and semi-arid areas the natural recharge by river is frequent; the sources



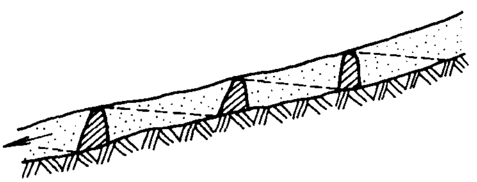
4.9 Surface infiltration of flood water; Pampa del Tamarugal, Chile. After Fritz et al.

are usually intermittent and ephemeral streams (when a temporary river water level is above the groundwater level), if any. The régime of perennial streams during dry periods is formed by the baseflow, and the contribution to the groundwater sources occurs usually during flood periods. A continuous recharge is typical for man-made canals (Smiles, Knight [23]). A seasonal recharge of groundwater from the rivers was described by Fritz et al. [24] in Parma del Tamarugal, where exogenous water resources originate in the upper regions of the Andes and water for recharge comes from the surface infiltration of flood water through creek channels (Fig. 4.9).

Water from ephemeral stream of the wadi type forms locally important sources of groundwater in parts of North Africa, Sudan etc. (Besbes et al. [25]). A distributed parameter wadi-aquifer model was developed by Illangasekare et al. [26].

Locally important sources of water which infiltrate under the sandy bottom of the river exist in Gwembe Valley, Zambia. The inhabitants traditionally mistrusted the water in wells and during periods when the rivers dried up, they dug the water from sandy beds down to the impervious rock. A simple underground dam (Fig. 4.10) or system of dams contribute to the stabilisation of the underground water storage.

Important sources of groundwater recharge from rivers are found in areas where ephemeral rivers terminate in depressions known in Africa as shebkas or shotts.



4.10 Groundwater conservation in the sandy river beds of intermittent rivers.

Here the topography of the region plays a dominant rôle, and thus through a diversion of the river channels the recharge can be at least partly regulated. The infiltration from the intermittent lakes which form in the depressions is usually delayed. This results in water loss caused by evaporation from the surface of such lakes.

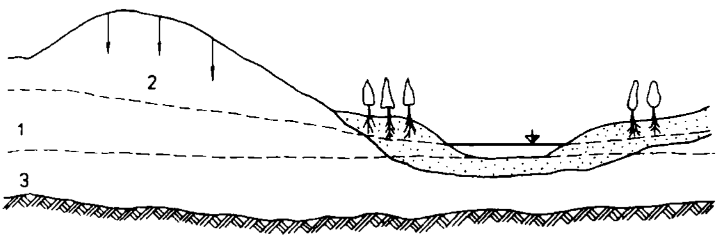
Winter [28] described the process of recharge from lakes in temperate regions. He described the stagnation point, which is located on the lake divide where the head is at the minimum. If the head at the stagnation point is higher than the lake level, water cannot flow from the lake to the headwater.

Steenhuis et al. [29] analysed groundwater recharge on eastern Long Island in the area where a more or less constant recharge rate (50% of the annual precipitation) was expected, and found that the recharge was seasonally variable. About 75 – 90% of precipitation between October and May contributed to the recharge in that area.

The seasonal recharge pattern can be modified by an artificial recharge. The methods are discussed in Chapter 7.

4.3 GROUNDWATER STORAGE

Several types of stored groundwater are recognised in the aquifer (Fig. 4.11). For the so called live or active storage, an interaction between groundwater and baseflow in the stream network is typical. Once the live storage has been exhausted the streams dry out, unless supplied from the subsurface flow and flood water. From another point of view, live storage is considered as an intermediary between recharge and baseflow. The recharge is usually considered as a direct product of precipitation within the hydrological or hydrogeological boundaries of the region involved. However, live storage also can be supplied from other sources located outside the boundaries. Such types of groundwater sources are called exogenous. In arid or semi-arid regions the exogenous sources can originate from humid or semi-humid areas, or the gradient of the exploited groundwater system may be responsible for the long distance transport of exogenous sources to another place.



4.11 Live storage – *1*, groundwater recharge – *2*, and dead storage – *3*.

Localised exogenous sources are found in those areas where a sharp difference in relief and rainfall exists over a short distance. Some valleys are fed by rather small aquifers from mountains in their vicinity.

From the hydrological point of view, the interaction between live storage and the baseflow in characterised by the recession curve.

The live storage also can be depleted artificially by pumping, or replenished by artificial recharge. For inland areas, live storage is located above the outlets, and in coastal areas it is above the border line with sea saltwater.

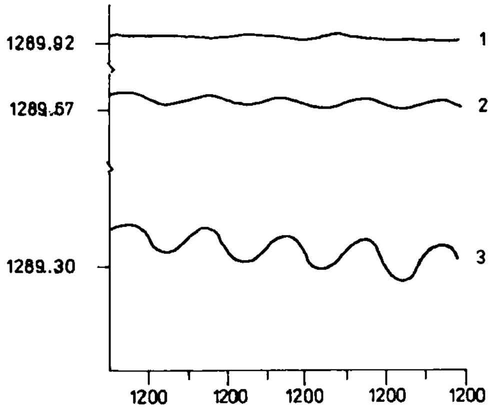
A simple formula characterises the interaction between live storage and baseflow:

$$G = -\frac{Q_g 3.6}{A \ln K} \text{ mm} \qquad 4.7$$

where Q_g is the instant baseflow, in $m^3\,s^{-1}$; A is the drainage area in km^2; K is the recession coefficient for hourly intervals, and G is the live storage, in mm. The last value is calculated as a mean for the area A; we can expect that it will be unevenly distributed. Some parts of the drainage area may not contain any live storage at all.

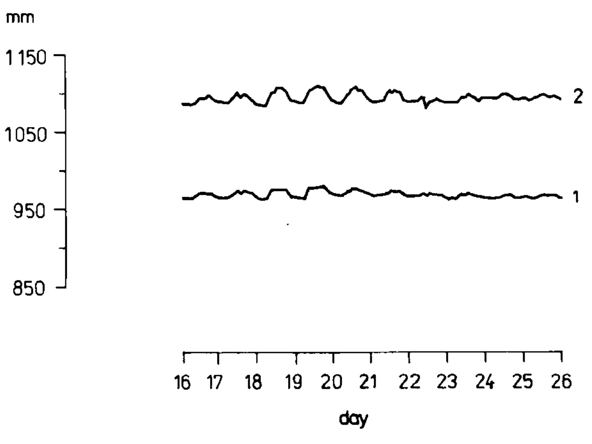
In contrast with live storage, the dead storage is normally stagnant. As soon as it is reduced by pumping, however, a continuous flux in the aquifer may occur, and replenishment even from exogenous sources may be initiated. Thus when the pumping of dead storage terminates, it may be replenished by a recharge from precipitation or from exogenous sources.

Both live and dead storage can also be reduced by evapotranspiration, provided the root system is so developed that it can tap water through the zone of capillary rise. Tromble [30] studied the water uptake of mesquite trees in a semi-desert region where the runoff from the catchment headwaters was absorbed by ephemeral streambeds, and taken from there by riparian vegetation. A peculiar pattern of groundwater level under such conditions is seen in Fig. 4.12. In a temperate region, a similar effect was observed upon groundwater level fluctuation of the live storage which



4.12 Groundwater consumption by mesquite trees. March – *1*, April – *2*, May – *3*. Arid region, U.S.A.; after Tromble.

supplies the baseflow of Plešivec creek in northern Czechoslovakia (Fig. 4.13). In some tropical regions 95% of groundwater storage, live and/or dead, is consumed by evapotranspiration.

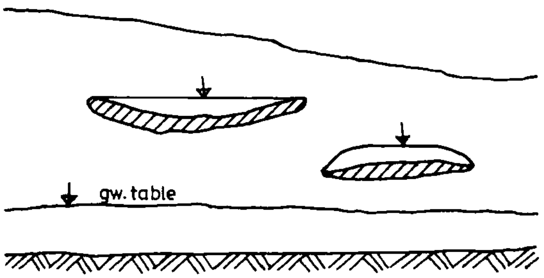


4.13 Daily fluctuation of groundwater level in the summer months, Plešivec creek basin, northern Czechoslovakia. Garczynski explains this phenomenon by electro-osmosis [32].

A special feature of some aquifers is what is called irreversible storage. Because of the special physical properties of the aquifer, when groundwater level is lowered the sediments undergo an irreversible compaction and the interstices cannot be repeatedly refilled by water.

Some hydrologists recognise as a special type of groundwater storage the so-called local reserves, located in the vicinity of pumped boreholes in the form of depression cones of various size. After a prolonged period of pumping these cones may or may not spread over extensive areas, this depending on the geological structure of the aquifer.

Perched aquifers are another type of groundwater reserve. They are formed within the main aquifer when groundwater reserves of relatively small volume are separated from the main body of groundwater by an impermeable stratum of small extent (Fig. 4.14).



4.14 Groundwater perched above impermeable strata.

4.4 SAFE YIELD, GROUNDWATER EXPLOITATION AND GROUNDWATER MINING

Safe yield is defined as the amount of water which can be taken from the aquifer indefinitely without producing an undesirable result. An undesirable result is often defined artificially. From a hydrological point of view, the maximum safe yield is equal to the long-term mean annual recharge of any origin or, in other words, to the amount of water which under normal circumstances leaves the basin as a natural baseflow. From a practical point of view, however, it has to be taken into account that the hydrograph during the dry periods is formed by baseflow only, and thus any lowering of the baseflow may produce undesired results in the water economy downstreams.

Exploitable groundwater resources are closely linked with safe yield and with the natural recharge from precipitation and from exogenous sources. We can recognise:

1. potentially exploitable groundwater resources which represent a maximum close to the live storage, and are limited only by the technical problem of setting an adequate system of boreholes to utilise the groundwater to the maximum; and,

2. actually exploitable groundwater resources governed by technical, environmental and legal requirements on the minimum baseflow and/or minimum groundwater level.

A difference between groundwater exploitation and groundwater mining is related to the difference between renewable and non-renewable groundwater resources. Safe yield, called also sustainable yield, is sometimes considered as the upper limit of that exploitation which will not result in the formation of what are called overdraft areas. In another interpretation, undesirable results occur when the groundwater storage cannot be replenished by a natural recharge in a reasonable period of time, or when a prolonged abstraction results in the intrusion of saline water, or in deterioration of water quality.

Thus a permanent long-term abstraction of water in excess of calculated natural recharge, and in excess of live storage, can be considered as groundwater mining. However, if in some aquifer the volume of water which is stored as dead storage is many times the volume of the live storage, then a temporary use of water exceeding the natural recharge can be allowed without any damage, provided that a sufficient period without any removal will be left for water recovery. Unfortunately, usually once such a type of water removal has been initiated, the demand steadily increases. Currently, a heavy drop in piezometric levels as a result of groundwater mining is reported from many parts of the world.

Walton [31] concluded: "From the water users' point of view, the so-called sustained exploitation means that the water supply is being frozen at a certain level because of the seemingly unfounded apprehensions of a hydrologist". This is often supported by the fact that almost all the adverse effects of groundwater removal are brought about by a slow process. Walton suggests that the sustained yield of an aquifer should be estimated by the following steps:

1. Determine the average annual replenishment.
2. Identify the most stringent constraint.
3. Find the quantitative relation between water level elevation and the occurrence of this unacceptable effect.
4. Define minimum water levels for the whole aquifer for the above-mentioned key positions.
5. Compute the rate of natural outflow that will occur when a quasi-steady state of flow, commensurate with minimum water levels, is established.
6. The sustained yield is the difference between 1. and 5.

A probability concept in safe yield analysis is discussed in Section 8.2.

Considering the great age of groundwater which in some regions interacts with the stream network, the question may arise as to whether the water balance approach as discussed in Chapters 2 and 3 is generally applicable. It is true that a water balance equation can be set up for any region; however, in some cases precipitation exerts very little influence on the immediate formation of the groundwater component and its seasonal fluctuation. This occurs particularly in large subsurface reservoirs where the precipitation is transformed into the discharge through a long distance communication and/or piston effect.

As an example, we can examine the water blance equation valid for two neighbouring basins B_1 and B_2:

Precipitation mm	Surface runoff mm	Baseflow mm	Loss mm
680	42	160	478

Let us suppose that the area of each of the two basins is equal to 60 km^2; however, the hydrograph analysis indicates the recession coefficient $K_1 = 0.998$ and $K_2 =$ $= 0.999\,996$, respectively. Obviously, the mean groundwater outflow of both basins is equal, supposing the same groundwater outflow of 160 mm in each basin:

$$Q_g = \frac{160 \cdot 60 \cdot 1000}{3600 \cdot 24 \cdot 365} = 0.304 \text{ m}^3 \text{ s}^{-1}$$

However, the corresponding mean live storage in basin B_1 is only:

$$G_1 = \frac{-0.304 \cdot 3.6}{60 \cdot \ln 0.998} = 9.11 \text{ mm},$$

while in the basin B_2

$$G_2 = \frac{-0.304 \cdot 3.6}{60 \cdot \ln 0.999\,996} = 4599.99 \text{ mm}.$$

Accordingly, in basin B_1 the groundwater outflow component will fluctuate much more than in basin B_2, and the contribution of precipitation (partly transformed into groundwater recharge) will have a direct effect on the seasonal fluctuation of the groundwater outflow.

We can conclude that the water balance approach is always closely linked with the assessment of the recession coefficient, live storage, and active drainage area which are all pertinent to the water balance equation. Therefore hydrograph analysis should be always provided within the framework of a water balance calculation.

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5 THE SYSTEM APPROACH TO GROUNDWATER RESOURCES ASSESSMENT

Any analytical study which helps to identify and select a preferential course of action from several feasible alternatives is called system analysis, if it is based on a logical, systematic approach in which assumptions, objectives and criteria are clearly defined and specified. Through the system approach the decision maker can broaden his information base, achieve a better understanding of interaction between the subsystems, and assess the consequences of several alternative courses of action. Also, the optimum course of action can be chosen to accomplish the specified results.

In a model approach, system analysis is defined as a problem-solving technique based on building a replica of a real-life system or situation. By experimenting with replicas one can obtain some insight into real-life problems. Such a system is often defined by a series of mathematical formulae and expressions. Because the parameters which characterise the system and the factors which influence the parameters are important in the model approach, a thorough knowledge and understanding of the system to be analysed is essential. In water resources management and in hydrology many factors affecting the system are unknown, or cannot be quantified with the necessary accuracy, and thus the relevant models do not describe the actual system exactly. Therefore the evaluation of the proximity of simulation should also become part of the system approach.

In general, system analysis employs methods and techniques which, in a simplified form, are accessible to many persons with some technical background, and reasonable results can be obtained by other than computerised model methods; however, for modelling complicated systems, better results can be expected from computer-oriented methods.

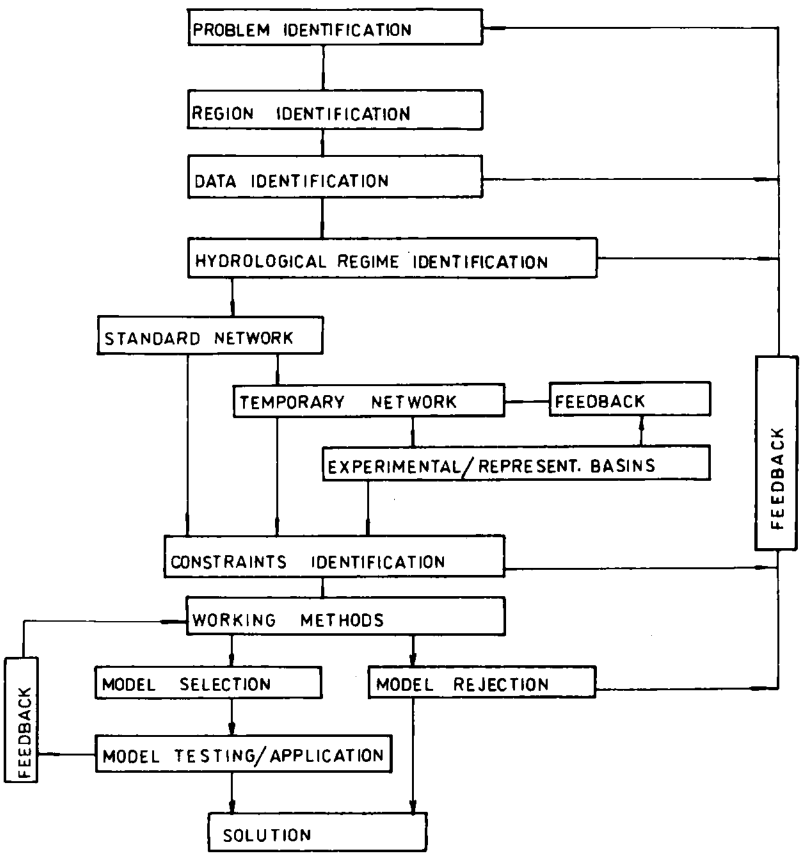
A water resources system analysis should pass through five stages (Biswas [21]):

1. Identification and explicit statement of objectives.
2. Translation of objectives into measurable criteria.
3. Identification of alternative courses of action which will satisfy the criteria.
4. Determination of consequences that follow from each alternative.
5. Comparative evaluation of the consequences of the alternatives in terms of criteria.

Many modifications of the water resources analysis can be set up depending on the type of problem to be analysed. For groundwater resources analysis, an alternative to the system approach is given in this Chapter.

5.1 IDENTIFICATION OF THE PROBLEM

A proper identification of the problem to be analysed and/or modelled is the result of initial discussions between the users/investors on the one hand and the person or team responsible for the solution on the other. This may appear to be the easiest part of the project to accomplish; however, a clear formulation of the problem, a preliminary approval of the working methods proposed and the sources of information available at the beginning and in the subsequent stages of the projects, should be set up as the very first step. Often the users who finance, and sometimes even supervise, the general project may have a limited knowledge of the methods proposed as an optimal solution and may request an inadequate or obsolete type of solution which is, however, closer to their experience. Then changes and additional work as a result of incomplete identification of the problem in the earliest stage may delay the latter phases of the project; delays for which the analyst will be blamed.



5.1 How to solve the problem through the system approach.

Therefore a simple contract, with a clear list of specifications of requested results and approved methods, is highly recommended. When a model solution is included, the right to adapt or reject it should be included in the contract, to cover the possibility that in further identification the data will be found inadequate for the model solution originally suggested.

In general, a simple method of solution should be favoured by the analyst, particularly when he feels that the project users will not be sufficiently competent to benefit from an advanced approach. The degree of complexity should be chosen also according to the importance of the problem, and in any case there is no need to apply very advanced methods when the data available and means of further identification are only very approximate.

Fig. 5.1 illustrates how the identification of the problem can be resolved through the feedback based on information obtained at any stage of the identification.

5.2 IDENTIFICATION OF THE REGION INVOLVED

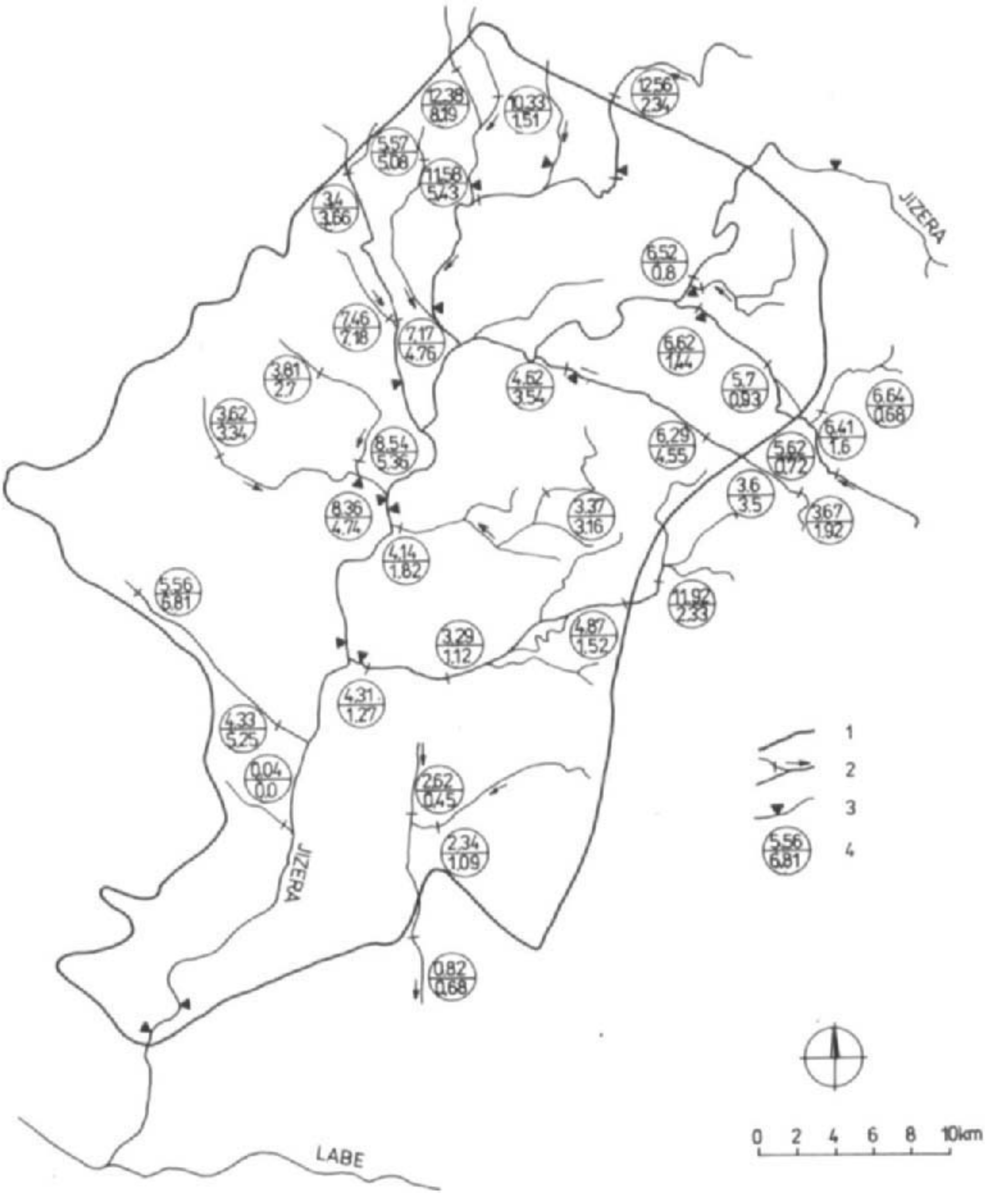
Field surveys of the ungauged areas and gauged basins are the main source of information. A survey of reports, studies and papers concerned with the hydrology and water resources management of the region will provide additional information. Last but not least, all available maps and charts related to the regional hydrology must be collected and analysed.

As in many hydrological studies, the region involved is not always limited by hydrological divides or by hydrogeological boundaries. In some cases, the definite limits of the region have to be partly or completely identified at a later stage of the project. The analyst must always be prepared for the contingency that the boundaries may be changed several times during various phases of the project. It is best to set up a general solution which is independent of the size of the area.

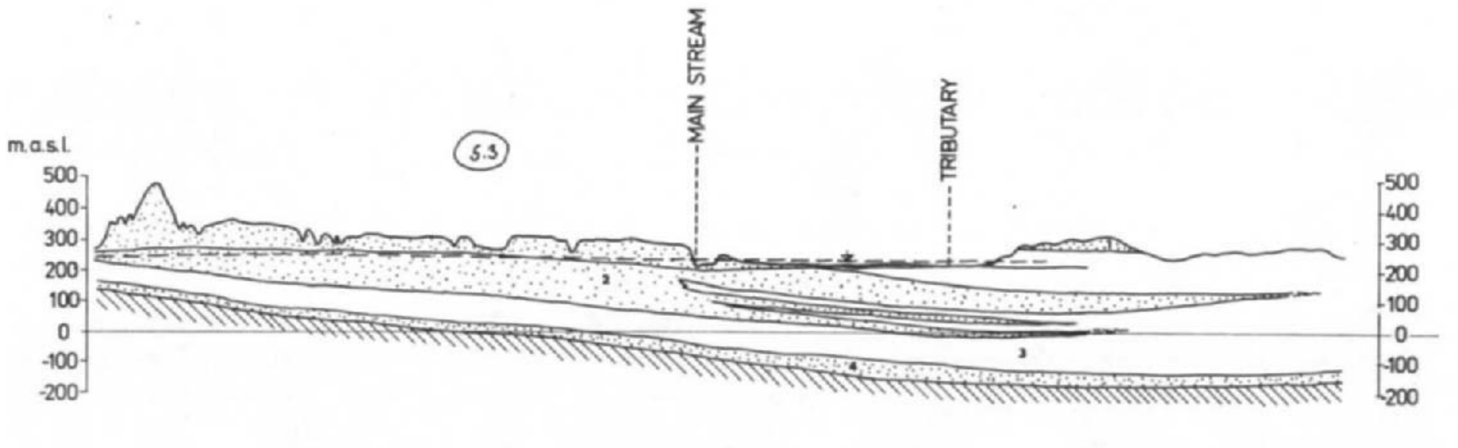
To identify the region, good quality maps, charts and air photographs are essential. Also two- and three-dimensional graphs derived from the data, satellite images, sectional views, diagrams, and sketches all provide additional and often valuable information.

From the many maps commonly available, those used will contain the features relevant to the hydrology and hydrogeology. Some hydrologists prefer to prepare their own compilation from several maps and their own observations. Even if such a working chart may never be published, in the initial stage of the project it can be an efficient tool. As an example, in Fig. 5.2 a chart is given of the discharge measurements, tracing the development of the baseflow in the region. Here the stream network and low flow discharges/yields at selected river cross-sections provide essential information.

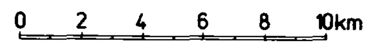
Similar charts can be compiled from geological and hydrogeological maps; the hydrological divides, stream network, afforestation, urbanisation, etc. are obtained



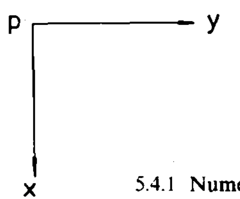
5.2 Working chart of instant baseflow yields ($l\ s^{-1}\ km^{-2}$), based on two field measurements. Regional boundaries – *1*; stream network and measured cross-sections – *2*; gauging stations – *3*; yield for first/second measurement – *4*.



5.3 Cross-sectional sketch of sedimentary formations in a Cretaceous region. Coniacian formations – *1*; live and dead storage in Upper Turonian formations – *2*; aquiclude – *3*; Cenomanian water-bearing formations – *4*.



	330	331	332	333	334	335	336	337	338	339	340	341	342	343	344	345	346	347	348	349	350	351	352	353	354	355	356	357	358	359
290	181	198	206	208	230	226	218	186	132	117	112	128	126	137	117	121	133	143	139	145	159	150	135	142	149	147	151	162	189	219
291	154	176	189	197	213	218	222	206	175	134	116	120	122	135	117	109	123	136	140	145	145	160	148	146	151	152	157	157	193	214
292	179	171	190	209	213	206	224	204	179	152	131	118	120	119	122	116	116	128	134	160	151	154	135	141	153	160	163	165	180	206
293	189	178	198	198	202	210	231	208	188	157	119	109	111	112	117	117	113	119	121	148	150	132	120	123	151	168	168	161	170	188
294	181	190	159	178	177	182	207	223	218	192	143	118	117	116	111	118	118	106	110	132	140	118	110	123	134	144	131	148	146	157
295	175	186	156	166	174	177	178	198	198	199	185	145	125	117	119	127	128	108	120	122	134	111	100	103	115	125	126	131	128	120
296	181	150	145	141	146	121	123	143	144	148	151	154	125	123	116	120	123	113	117	114	116	100	90	95	110	114	120	135	126	106
297	151	122	122	118	116	101	102	110	120	129	130	132	132	115	122	122	123	111	114	117	120	111	111	106	107	112	128	130	129	99
298	132	111	111	110	98	91	100	114	124	135	127	125	140	132	119	117	111	100	101	103	112	110	121	106	115	128	123	114	104	95
299	122	112	96	93	92	94	104	121	137	151	132	126	117	108	106	106	101	93	94	95	107	101	98	112	124	123	113	98	95	97
300	118	108	97	97	101	107	117	136	141	151	139	118	108	97	103	117	109	100	95	102	105	103	102	111	123	119	110	107	104	108
301	105	105	103	98	108	111	128	130	151	160	144	115	103	94	101	108	105	95	103	106	114	109	103	102	105	120	131	133	117	105
302	93	105	106	96	101	102	112	121	127	149	131	103	88	81	88	89	88	93	118	118	113	98	93	89	94	117	141	146	130	101
303	103	117	109	100	91	98	104	105	124	148	115	100	90	91	91	91	88	88	105	109	101	98	102	93	92	105	120	140	132	119
304	111	117	102	90	95	96	102	109	118	149	120	93	92	94	95	92	85	88	93	97	95	101	115	99	87	91	98	138	157	139
305	114	112	99	92	95	92	92	105	115	131	104	88	85	87	95	111	104	100	94	102	100	104	113	111	100	97	115	157	172	154
306	118	106	94	89	100	89	86	96	112	127	100	85	83	85	90	115	145	140	111	113	105	93	100	107	103	109	129	166	167	162
307	97	95	86	86	95	94	91	96	101	109	99	90	86	86	92	101	117	126	117	113	106	98	96	98	99	106	115	131	121	136
308	91	97	94	81	86	92	104	111	111	109	93	94	90	84	91	98	99	98	102	97	91	95	94	98	91	95	103	106	120	133
309	92	96	104	101	93	98	102	119	135	107	97	100	93	104	105	91	93	95	99	92	92	99	94	102	96	92	95	101	124	157
310	98	107	109	100	95	99	92	89	97	104	100	99	102	109	99	90	91	101	94	91	94	94	92	109	104	102	97	100	110	134
311	91	117	124	106	95	90	88	88	95	95	112	108	92	93	95	93	89	101	89	85	91	96	103	111	118	108	101	101	101	101
312	99	115	126	110	98	97	90	92	91	94	122	119	89	95	102	88	88	100	87	90	89	92	98	111	119	112	99	104	109	99
313	101	110	117	113	111	100	89	83	81	87	109	114	97	97	106	94	88	87	85	87	90	90	98	103	104	110	95	100	110	103
314	106	117	111	108	111	98	83	79	83	85	89	98	90	97	93	91	84	86	91	89	92	91	96	101	101	98	105	104	100	95
315	113	117	110	104	96	93	89	89	92	87	85	87	86	96	94	92	93	92	101	94	92	90	93	90	87	93	97	94	93	94
316	120	114	115	107	106	100	93	90	87	91	87	85	91	101	97	94	99	90	96	93	95	91	84	87	87	89	92	89	90	89
317	106	107	120	122	122	132	112	95	93	94	100	95	95	96	96	96	94	92	93	102	101	91	86	95	98	105	93	91	94	98
318	95	108	127	136	151	141	132	127	107	107	112	108	102	106	100	98	96	85	90	91	97	91	88	97	110	106	102	95	104	122
319	97	121	136	141	138	137	136	137	117	113	116	118	105	122	125	102	96	90	88	90	88	89	89	99	105	98	99	109	105	120
320	101	113	117	130	123	110	119	109	114	123	115	103	114	120	126	103	99	94	88	88	94	96	91	99	113	98	92	98	112	133
321	100	104	106	109	100	93	97	95	96	107	105	95	103	99	103	97	92	89	88	94	92	93	92	96	100	95	89	102	125	152
322	102	99	101	100	93	90	90	89	85	93	93	93	95	96	99	94	90	92	96	94	88	90	89	91	96	90	97	103	139	164
323	96	98	104	95	91	82	85	91	89	89	93	93	92	96	100	94	82	85	91	94	89	90	88	88	94	94	105	127	155	154
324	108	109	113	100	93	91	88	92	91	92	90	96	100	90	95	92	86	88	94	93	93	90	89	92	98	111	135	151	170	167
325	112	111	116	120	116	98	93	95	94	91	92	93	97	97	97	94	93	90	89	93	97	94	97	101	106	110	124	121	149	156



5.4.1 Numerical two-dimensional chart of the groundwater level fluctuation in a small area.

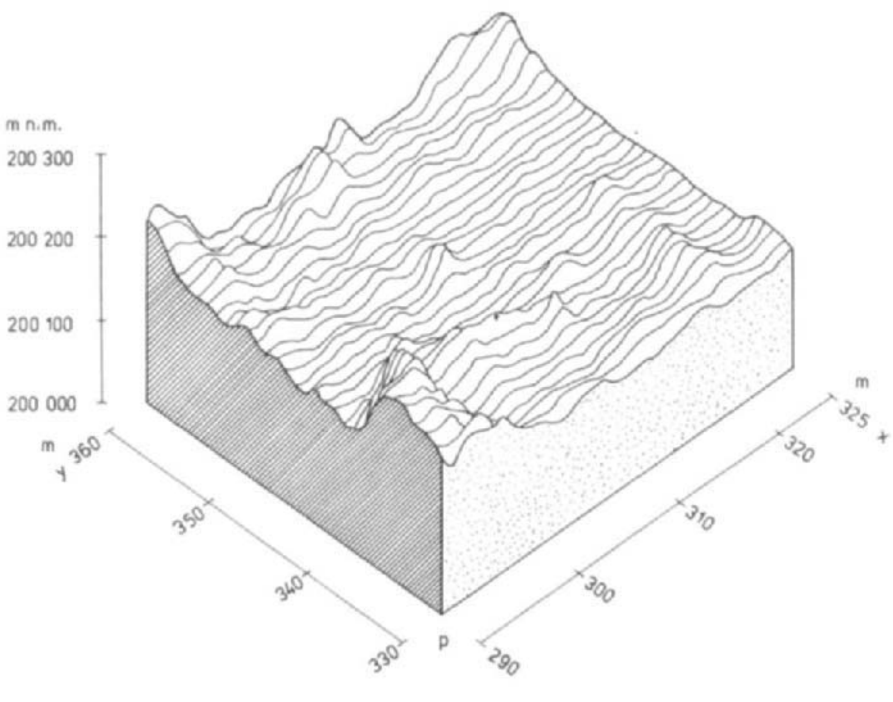
from standard topographical maps. Other sources of information are soil maps, land use and morphology maps, and various ecological maps. A scale of 1 : 50 000 is often found to be the most convenient. Sometimes, however, this may be too large when an extensive region is studied. On the other hand, maps on a larger scale are preferable for detailed analysis of small areas.

Additional features such as the water quality and characteristics of the aquifers can be plotted on special maps. An unreasonable amount of data on a single map cannot be recommended.

Cross-sectional diagrams are drawn when a vertically complicated system is analysed (Fig. 5.3).

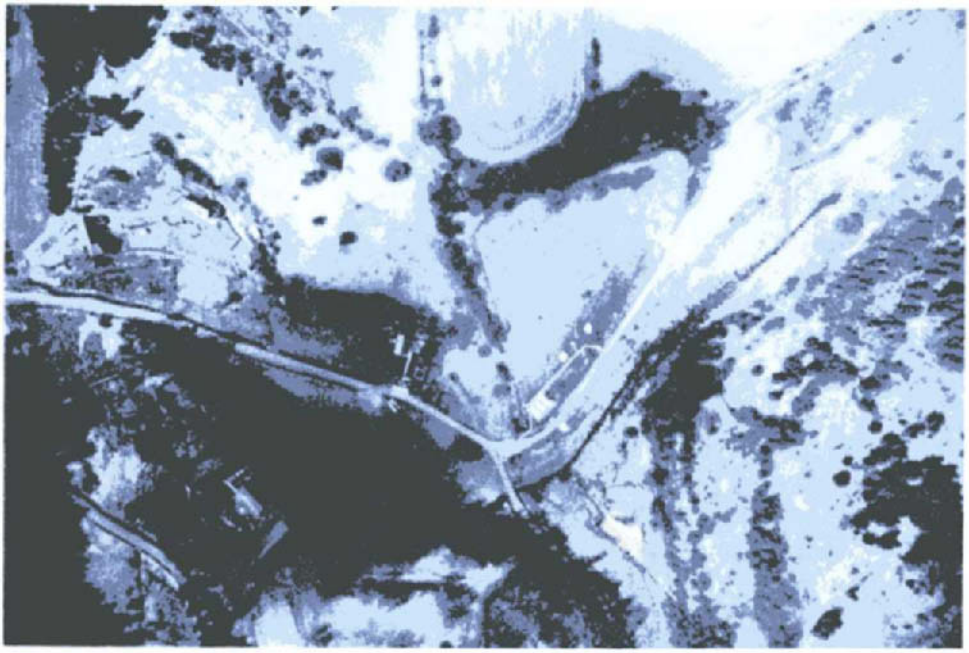
A great amount of data, often stored in the database of a particular project, can be used to plot special maps emphasizing particular features relevant to the project. They are produced by computer peripherals. Fig. 5.4.1 shows a numerical chart of levels, which serves as a guide for an initial orientation in the project area. In Fig. 5.4.2 is the same chart, this time drawn on a three-dimensional scale. A comparison of these charts with a topographical map of the area facilitates the identification.

Aerial photograps taken on a conventional film are often used as a source of additional information. They record anything visible on the ground, and thus it might be seen that from a hydrogeological point of view they have a limited use.



5.4.2 Three-dimensional chart of the groundwater level fluctuation in the same area as in Fig. 5.4.1.

However, from the visible surface feature we can often make an assessment of the hydrogeological characteristics of the studied area. On standard monochrome photographs, wet ground appears darker than dry ground, muddy water is lighter than clear water, and deep water is darker than shallow. A stereoscope can be used to give stereo-viewing. In a strictly topographical approach the distortion of scale and shape has to be taken into account; in a standard hydrological approach there is rarely need to be concerned with the distortion.

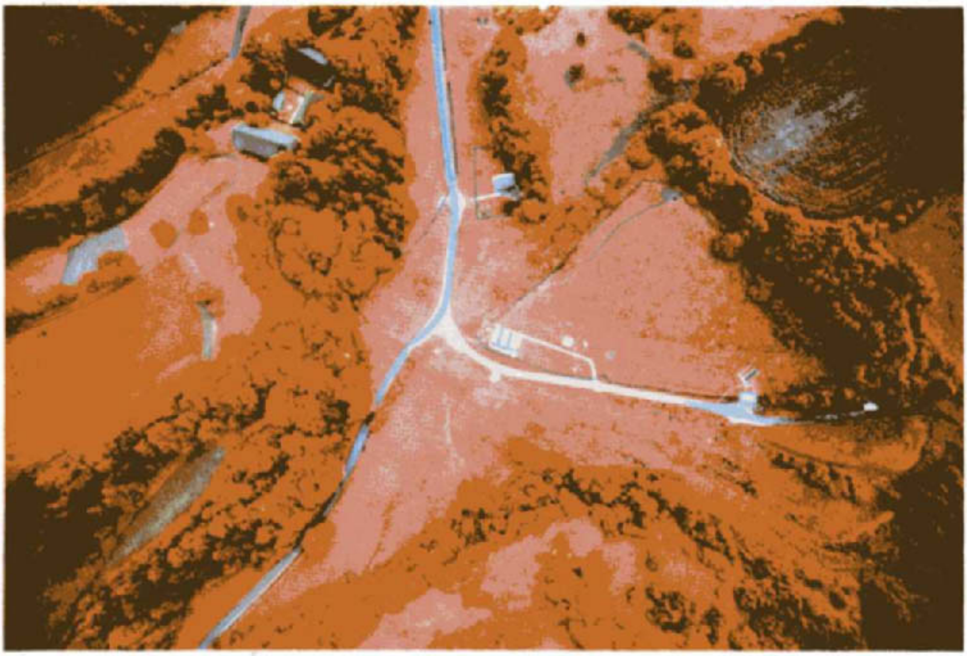


5.5 False colour image of the experimental area at the end of the non-growing season. From in situ measurements no transpiration was observed on the lysimeter and on the trees. Groundwater recharge was still very high.

With the development of remote sensing methods, non-conventional films have become widespread. A ground reconnaissance is always recommended before any definitive conclusion is made.

In Fig. 5.5 and 5.6 is an example of the use of infrared film in water balance analysis. Here the area photographed was one in which experimental plots are situated, with ground observation points measuring transpiration. The blue appearance of the area is typical of the non-growing season, when an intensive groundwater recharge occurs. During the high transpiration period, with no groundwater recharge at all, a red image prevails. Through a comparison of the optical densities of blue, green and near infrared components in each experimental plot, the results can be extrapolated to the whole area (See 2.8.7).

Further examples relevant to various hydrogeological and geological phenomena can be found in specialised literature [1].



5.6 False colour image of the same experimental area in the middle of the growing season. A full transpiration rate of 2.1 mm day^{-1} was observed on the lysimeter, and 2.5 mm day^{-1} on the trees. Soil moisture deficit was about 25 mm.

It is often believed that the geology of the area is a decisive factor in demarcating the region in which hydrogeological analysis is to be performed. In other words, it is expected that on a uniform geological structure a uniform water balance and groundwater behaviour will be found. This is far from being true. A comparison of two charts will serve as an example, one of them indicating the simple geological features of the studied area, the other giving additional information on its morphological structure (Fig. 5.7 and Fig. 5.8). Here a fast subsurface flow can be expected in the steeper parts of the region, while the recharge areas supplying the deep circulation are limited to the flat parts or to the areas in which the so-called preferred pathways can be identified.

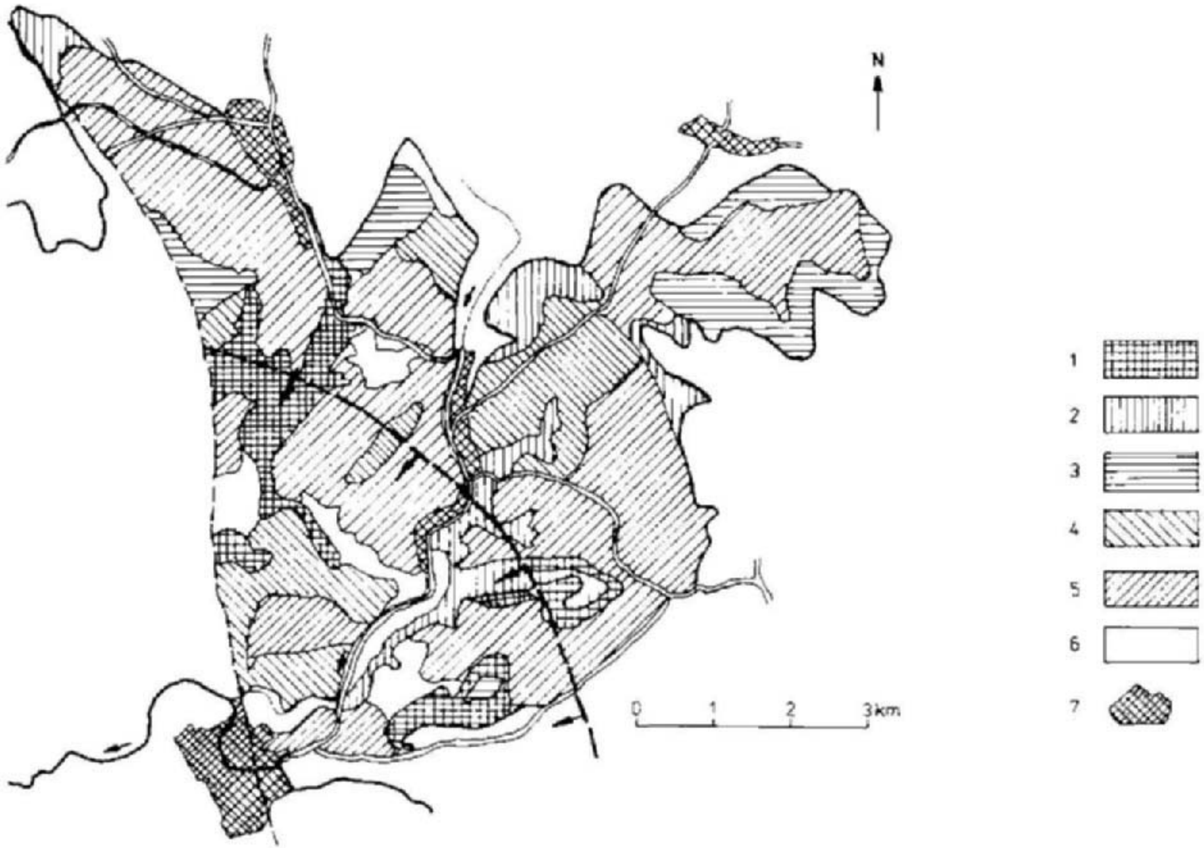
5.7 Region consisting of area suitable for groundwater recharge — *1*, and impervious formations — *2* under which an artesian groundwater storage is formed.

Any survey of reports and publication relevant to the problem must also include the study of already existing maps and charts. As with the reports, it is not recommended to use such sources without careful control procedures, otherwise the new results may be undesirably biased by the earlier errors.

5.3 IDENTIFICATION OF DATA

The main sources of information are records published in yearbooks and special reports. Highly valuable material can be found in the archives of organisations concerned with meteorological, hydrological and hydrogeological surveys; however, frequently the data are unprocessed and may contain many gaps. Data of such type need to be checked and corrected; in the initial stage we have to decide whether these data are worthwhile using further, because additional extensive corrections and processing may considerably increase the budget.

Long-term precipitation records are usually available, although occasional prolonged gaps may considerably reduce the original length of the records.



5.8 The same region as in Fig. 5.7. Plotted morphological features indicate the steep part of the potential recharge area in which subsurface lateral flow prevails. Thus the active recharge are shrinks to a fraction of the potential recharge area. Slope >20% – *1*; 16–20% – *2*; 12–16% – *3*; 8–12% – *4*; 4–8% – *5*; <4% – *6*; urbanised areas – 7.

Observations of the river stages/discharges are usually shorter, and prolonged periods of missing records are frequent.

In the initial stage of the analysis a time period of records should be chosen which includes a maximum number of observing points, and which can be considered representative when compared with the long-term mean and/or totals. Sometimes it is necessary to use data from a period of time which cannot be regarded as representative. In such a case, a thought must be paid as to how the obtained results will be linked to the representative period. We have to bear in mind that in some situations a linear relation based on the comparison of long-term and short-term rainfall records may not be entirely correct.

The form in which data are stored is also important. The most convenient technique would be to have the data available on magnetic tapes, discs or similar media which can be used for a given purpose without further processing. This, however, will happen only rarely. More often the data need to be reprocessed and gaps filled. Frequently it is found that some data come from areas affected by man's activity, and thus are not representative of the natural processes. Sometimes these data can be adapted when compared with the data from natural evironments. This can be done if the events are not too prolonged. The short-term effect of pumping on groundwater level fluctuation can be easily eliminated, but some difficulties can be expected in the correction of prolonged effects.

The consistency of data also has to be checked. Some data have been collected using different units, and the transfer periods almost always contain some errors. Also a change in the type of gauge or recorder may influence the consistency of the measurements.

An example of the data analysis approach is given in Fig. 5.9.

After checking the quality of data available, we have to decide whether some additional networks should be set up elsewhere in the region. It should be stressed that the data are not collected merely from the area identified at the initial stage of the project. Observation points outside the boundaries provide equally valuable information, because when using only the inside stations we have to extrapolate the data in the vicinity of the boundaries and this is always more difficult than the interpolation.

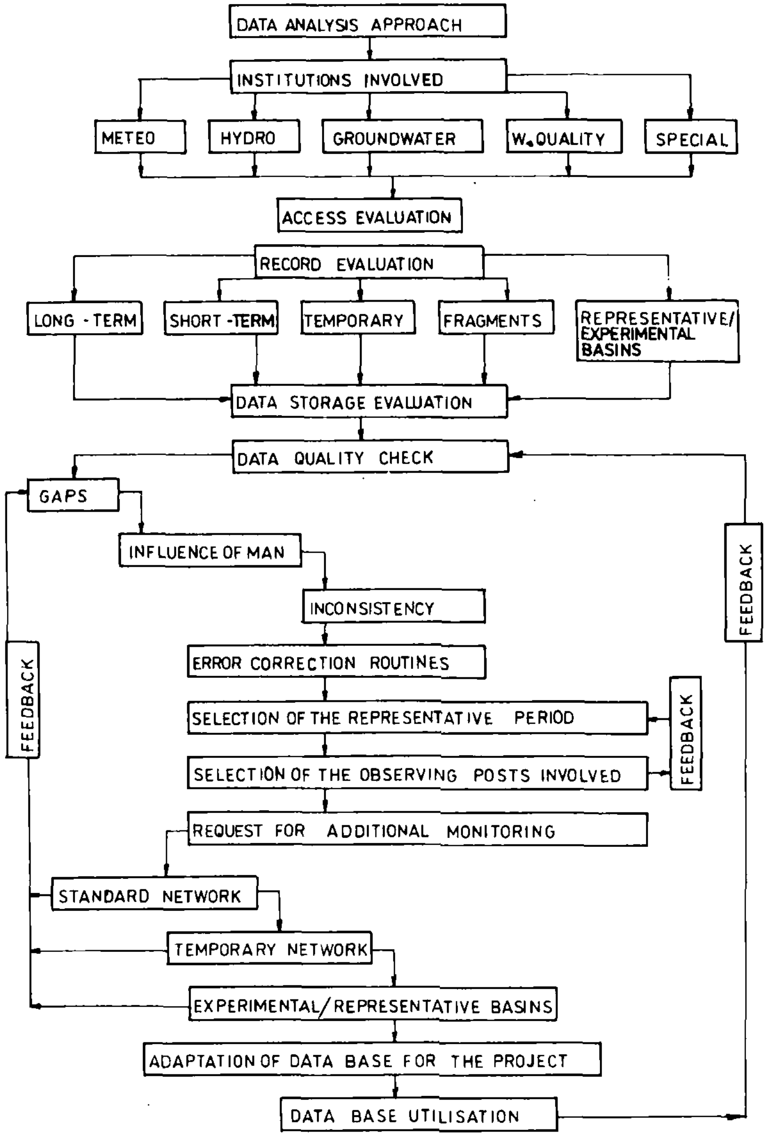
We often need to define a short period of observation available for a great number of observation points, and to determine whether the selected period can be considered representative. For this purpose we can define 95%, 99% or other confidence intervals using Student's table of distribution. In this manner we can estimate, within specified limits of confidence, the population mean μ. If for instance $-t_{.975}$ and $t_{.975}$ are the values of t for which 2.5% for the area lies in each "tail" of the distribution t, then a 95% confidence interval for t is

$$-t_{.975} < \frac{\bar{X} - \mu}{s} \sqrt{N - 1} < t_{.975}$$

In general, confidence limits for population means can be expressed as

$$\bar{X} \pm t_c \frac{s}{\sqrt{N-1}}$$

where the values $\pm t_c$, called critical values of confidence interval, depend on the level



5.9 Data analysis approach.

of confidence desired and the sample size N; $\bar{X}$ is the arithmetic mean; and s is the standard deviation of the sample.

In Tab. 5.1 we have the yield as observed in the 1974–1976 and 1931–1960 periods in a sedimentary structure of the Labe and Morava headwaters. We want to find out whether the 1974–1976 period is representative when compared with the 1931–1960 period. We determine

$$0.0140 \pm 1.94 \frac{0.001\,26}{\sqrt{7-1}} = \begin{cases} 0.0182 \\ 0.0098 \end{cases}$$

We can see that the mean yield for 1974–1976 lies within the confidence interval, and thus as far as the discharge régime is concerned the short period can be considered representative.

Tab. 5.1 Yield at some cross-sections in the upper Labe and upper Morava basins

Station	River		1974–1976 $l\,s^{-1}\,km^{-2}$	1931–1960 $l\,s^{-1}\,km^{-2}$
Nekoř	D. Orlice		0.0199	0.0198
Litice	D. Orlice		0.0172	0.0173
Slatina	Zdobnice		0.0191	0.0192
Sobkovice	T. Orlice		0.0148	0.0122
Machovice	T. Orlice		0.0130	0.0122
Albrechtice	M. Sázava		0.0117	0.0127
Mezihoří	Třebůvka		0.0048	0.0047
		$\bar{X}$	0.0144	0.0140
		s	0.00529	0.00526
		N	7	7

Sometimes we need to determine whether there is a significant difference between two means of random samples of different sizes N_1 and N_2. Assume that two random samples are drawn from normal populations whose standard deviations are equal. We suppose that these two samples have means and standard deviations given as $\bar{X}_1$, $\bar{X}_2$ and s_1, s_2 respectively. To test the hypothesis that the samples come from the same population, we use the t score given by

$$t = \frac{\bar{X}_1 - \bar{X}_2}{\sigma \sqrt{\dfrac{1}{N_1} + \dfrac{1}{N_2}}}$$

where

$$\sigma = \sqrt{\frac{N_1 s_1^2 + N_2 s_2^2}{N_1 + N_2 - 2}}$$

Again the distribution of t is Student's distribution with $N_1 + N_2 - 2$ degrees of freedom. As an example, let us look at the sample of mean annual precipitation from 1974 – 1976 and from 1931 – 1960 in the Labe headwaters. For the short period the size $N_1 = 17$ and for the long period the size $N_2 = 13$ are available (Tab. 5.2).

Using the above formulae we obtain

$$\sigma = 133.71$$

and

$$t = -0.31$$

Because for 28 degrees of freedom $t_{.95}$ is ± 1.70 and calculated t lies within those limits, we can conclude that the difference between the mean precipitation of each sample is not significant.

Tab. 5.2 Mean annual precipitation in 1974 – 1976 and 1931 – 1960 in the upper Labe basin.

Station No.	Name		1974 – 1976 mm	1931 – 1960 mm
024	Hořiněves		645.80	619
087	Holovousy		652.23	–
008	D. Králové		650.23	–
083	Jičín		609.93	637
005	Hostinné		743.10	715
007	D. Třemešná		676.20	646
009	Jaroměř		585.77	661
033	Doudleby		685.70	–
032	Slatina n. Z.		794.60	822
031	Zdobnice		1139.97	1105
051	Hroška		619.67	–
049	Dobruška		690.87	668
050	Č. Meziříčí		587.87	597
047	Albrechtice		646.03	663
015	Č. Skalice		639.93	703
016	Č. Kostelec		706.43	683
005	Hostinné		743.10	715
		N	17	13
		$\overline{X}$	695.13	710.31
		s	127.69	131.10

5.4 IDENTIFICATION OF THE HYDROLOGICAL RÉGIME

As the first step we have to finalise the identification of the hydrological and hydrogeological boundaries as related to the project area. It would be feasible to adapt the project boundaries so that they coincide with the hydrological or at least the hydrogeological divides: often, however, we have to accept that one or several streams may flow in the area or cross the boundaries several times; then the estimation of the water balance becomes more laborious.

5.4.1 Temporary monitoring network

To identify all of the possible features of the hydrological régime and set up a representative network, a temporary monitoring network has to be established. The number of precipitation and river gauges and observation boreholes depends on the type of project and budget available. For special projects even the density as recommended by WMO may not be found sufficient (Tab. 5.3).

Tab. 5.3 Minimum network density. After WMO

Region	Range for $\frac{\text{minimum}}{\text{provisional}}$ network, area km^2 per station		
	Precipitation	Evaporation	Hydrometry
Flat regions of temperate, mediterranean and tropical zones	$\frac{600-900}{900-3000}$	50 000	$\frac{1000-2500}{3000-10\,000}$
Mountainous regions of temperate, mediterranean and tropical zones	$\frac{100-250}{250-1000}$		$\frac{300-1000}{1000-5000}$
Small mountainous islands with very irregular precipitation, very dense stream network	25		140–300
Arid regions	1500–10 000	30 000	5000–20 000
Cold regions	1500–10 000	100 000	5000–20 000

The duration of the proposed operation of a temporary network depends on the type of project and the budget available. Normally such networks should operate for at least three hydrological years before some representative data can be obtained. A single complete hydrological year of observation should be considered as an absolute minimum. From the practical point of view the network should be put into operation well before the beginning of the hydrological year, because the instruments always need some adjustment and minor operational problems will have to be solved.

The types of instruments to be selected for temporary monitoring depend on the type of project. Unless there is enough experience and manpower, there is no need to install sophisticated instruments whose maintenance requires highly skilled technicians. Often simple standard instruments may be more convenient, and this also applies in terms of data processing.

Occasionally we may find that some stations and instruments which belong to what can be considered a permanent network need some maintenance and a routine check-up. Instead of complaining to the authority responsible for the network, it might be more feasible to do it within the framework of the project, particularly when the data can be utilised for this purpose.

While the existing network of rainfall and river gauges may be sufficient, more often than not special measurements will have to be started of, for example, the groundwater level fluctuations, soil moisture régime, and evapotranspiration régime. The main purpose of these observations is to specify the subregions in which the hydrological régime can be considered more or less uniform. Special measurements with portable instruments are recommended for this purpose; the installation of permanently recording instruments may considerably increase the total cost of the project.

Essential information on groundwater resources is based on the measurement of groundwater level. Advanced techniques have been introduced recently in this field; however, for many standard and special measurements, manually operated and simple instruments are recommended. The most common method of measurement involves a weighted and graduated tape suspended from a defined point. A chalk on the surface of the sensor attached to the end of the type is washed off at the groundwater level. In another modification, a cylinder open at the lower end and almost closed at the upper end gives a whistling sound when lowered rapidly to groundwater level.

The electric contact gauge is an instrument for the instantaneous measuring of water level and water temperature. As soon as the measuring probe touches the water surface, a signal lamp lights up on the cable drum. The distance of the water level from the defined point can be read on the measuring tape. The temperature reading is given by two signal lights and a knurled scale disc. The depth-measuring accuracy is usually less than one centimetre down to the depth of 100 metres, and slightly more down to the depth of 500 metres.

In principle, the same type of water-level recorders are used for the continuous monitoring of groundwater level as for river gauging. Strip chart water-level recorders, vertical or horizontal, are utilised. Sometimes vertical recording gauges are specially designed for the continuous recording of groundwater level. Their high sensitivity responds to the smaller changes of the water table. The box can be equipped with a casting which fits directly on the borehole, and no additional shelter is required. A clockwork mechanism can be supplied for a drum revolution period of up to 96 days.

A relatively new sensor for groundwater-level measurement is a pressure probe, also called a transducer. This sensor can be attached to a vertical water-level recorder equipped with an electric monitoring system. The water levels are recorded at adjustable time intervals. An interval operation is very economical, and ensures a long life-time for the battery.

Advanced types of such probes have an integrated amplifier to reduce the effect of fluctuating temperature. In fact, the probe can also be used for measuring the temperature. The narrow sensors are especially suitable for applications where the use of a float cannot be considered, particularly in boreholes of small diameter, at great depths, and in boreholes which are not exactly vertical. The probe and recorder can be installed separately. The power supply is usually provided by battery cells, which makes it easy to use the equipment in field.

The pressure probe can be attached to various types of data acquisition systems. The most advanced ones provide storage of analog or digital measuring signals on Eprom memory modules. Typical data acquisition systems with a rechargeable battery can be installed anywhere in the field. The data acquisition is initiated and controlled by a portable operating unit similar to a pocket calculator. Data are kept on semiconductor modules, and maintained for an unlimited period even in case of a power failure. For the interpretation of data stored on memory modules a reading unit is used, attachable to any computer. After the data have been transferred to the computer, the memory modules are erased by ultraviolet light and used in the field again.

For all other types of measurement, the instructions given in the WMO publication [2] are highly recommended.

Sometimes several pumping tests have to be performed in order to identify the hydrogeological régime. The boreholes selected for this purpose should not be viewed as part of the groundwater-level monitoring system. Short-term pumping tests serve usually for the assessment of the hydraulic properties of the aquifer, and they can be regarded as part of the identification of the hydrogeological régime; long-term pumping tests serve rather for the determination of the safe yield and thus they belong under the analytical part of the project.

Special attention must be paid to measuring all phenomena occurring during the extreme conditions, particularly during droughts. During these periodes additional special discharge measurements will contribute significantly to the identification of the hydrological régime. Fig. 5.2 depicts an example of the expeditional discharge measurements in a région with a permanent monitoring network of low density. Careful selection of the profiles in which the expeditional measurements are conducted, helps to determine the baseflow development of small streams and enables comparison of the baseflow from basin to basin. Beside the actual discharge values, we are also concerned with the determination of the baseflow yield from profile to profile. The measurements should be repeated several times during the fair weather periods, in order to obtain results which are statistically significant.

Another type of information which can be obtained through the expeditional discharge measurements is the identification of stream section with discharge deficits, where the baseflow returns temporarily or permanently into the aquifer. Faults crossing the river beds, karst phenomena in the vicinity of streams, and places with mining subsidence are typical causes of discharge deficits. An example of the results of special measurements under such conditions is given in Fig. 5.10.

Geophysical, geothermal and water-quality monitoring may contribute significantly to the process of hydrological régime identification. Geophysical and geothermal measurements, described in specialised literature, help towards the better understanding of the hydrological and hydrogeological structure of the region. The monitoring of chemical components (including radioactive identification methods) provides information on the mixing of water of different origins and its vertical and horizontal stratification. The formation of the baseflow from various aquifers, and their joint exploitation, can be traced by means of chemical and radioactive monitoring. Water quality analysis in general may serve as an indicator of man's impact on the groundwater régime.

Special working methods can be found in Unesco monograph [3]; as far as radioisotope methods are concerned the casebooks published by the International Atomic Energy Agency, Vienna, contain numerous examples of practical application.

5.4.2 Missing records

Gaps are often found in records, and are caused by the absence of an observer or as a consequence of defective instruments. In some cases the data can be completed by simple means, in others more advanced statistical methods have to be applied. When one daily reading of groundwater level is missing during a period of a stable or slightly fluctuating record, it can be filled by a simple interpolation between the values of the previous and the following day:

Month	Day	Observed level, cm	Correction
11	7	312	
	8	312	
	9	missing	312
	10	312	
	11	313	
	12	314	
	13	missing	315
	14	316	

In a more complicated case we can use a complete record from a well located nearby:

$$Y = \frac{(X_2 - X)Y_1 + (X - X_1)Y_2}{X_2 - X_1}$$

where Y is the missing record corresponding to the record X in the complete observation, and Y_2 and Y_1 are the values in the incomplete record which are known and correspond to X_2 and X_1 values in the complete record. An example is given as follows:

Month	Day	Complete record X	Incomplete record Y
4	26	315	120
	27	312	missing
	28	307	109
	29	306	105
	30	303	missing

For April 27 – interpolation:

$$Y = \frac{(307 - 312)\cdot 120 + (312 - 315)\cdot 109}{307 - 315} = 115.88 \doteq 116\,\text{cm}$$

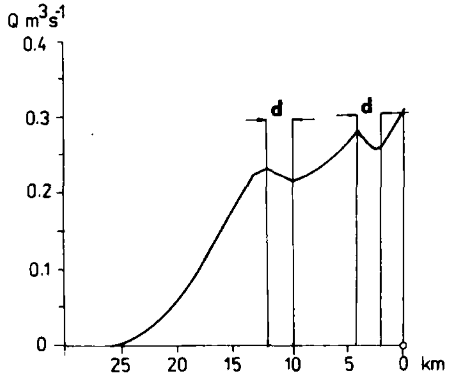
For April 30 – extrapolation:

$$Y = \frac{(306 - 303)\cdot 109 + (303 - 307)\cdot 105}{306 - 307} = 93\,\text{cm}$$

A more complicate situation occurs when, in the river gauge record, the peak flow of the hydrograph is missing; something which often happens to the recording instruments during high floods. Then the proper shape of the hydrograph can be plotted:

1. In accordance with the shape of the hydrograph at some previous similar event (similar amount of rainfall, similar gauge readings, etc.), or,
2. by a comparison of the incomplete hydrograph with the rainfall record, or,
3. by a comparison of the incomplete hydrograph with the hydrograph from the station upstream or downstream, and taking into account the time delay of the peaks.

Such extrapolation is less accurate than simple interpolation, and is sometimes entirely dependent on the judgment of the officer-in-charge.



5.10 Discharge development and discharge deficit – d in the longitudinal profile of a river; both supply additional information based on the expeditional discharge measurements.

When a longer period of records is missing, the statistical method of linear regression can be used. The aim is to set up a regression equation which describes the relationship between observed paired values, such as between discharge records upstream and downstream, Q_1 and Q_2:

$$Q_1 = a_1 Q_1 + a_0$$

where a_1 and a_0 are constants. Obviously, once such an equation has been set up for a period with complete records, it can be used to assess the missing records at both stations for a limited period of time.

It should be noted that some statistical relationships between various hydrological, hydrogeological and meteorological phenomena are non-linear. A preliminary assessment of the relationship can be made graphically.

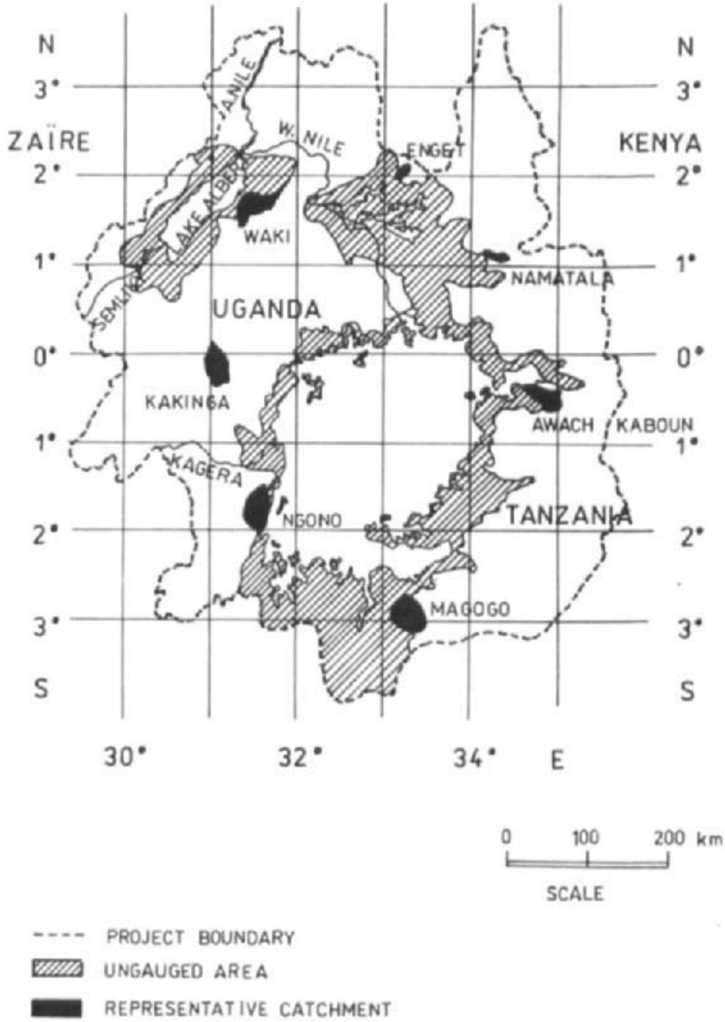
5.4.3 Experimental and representative basins

In some projects it is not feasible to establish an ideally dense network over the whole area. Financial problems, transport difficulties, and the unaccessibility of some areas do not allow the operation of evenly distributed observation posts. In such a situation, some intensively-observed areas can be established in the region. The areas selected should be representative for the region or for clearly defined parts of it. As a guide to assessing their representativeness, the type and range of climatic, vegetational, geomorphological, soil and geological characteristics of each selected area should be compared with those of its region or subregion.

It is convenient to define the area by basin boundaries, otherwise it may be difficult to generalize from the observed results. The basins established for the purpose of groundwater resources assessment are located on aquifers whose size may exceed by several times the drainage area of the basin. So the representativeness of the basin from the hydrogeological point of view should be clearly defined in the initial stage of observation.

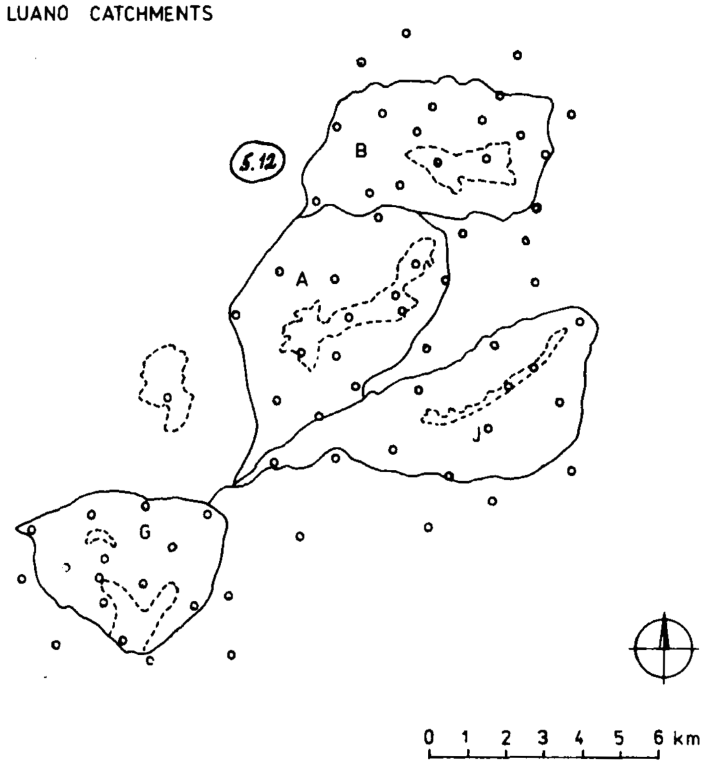
An example of representative basins which served for the calculation of the hydrological régime in large ungauged basins attached to the Great African Lakes is given in Fig. 5.11 [4]. The drainage area of some of these representative basins was several hundreds of square kilometres; in many cases studies of much smaller basins were found to be sufficient to the purpose. In general, the site of the representative basin depends on the purpose for which it is studied. When we need to collect a large amount of very specialised data for mathematical modelling, a smaller basin can be adequate. On the other hand, in order to study human impact and the influence of land-use changes on the hydrological régime, more extensive basins should be chosen in which local effects and deviations are eliminated.

Unlike those representative basins in which only natural changes of any kind are accepted, the so-called experimental basins are often subjected to induced



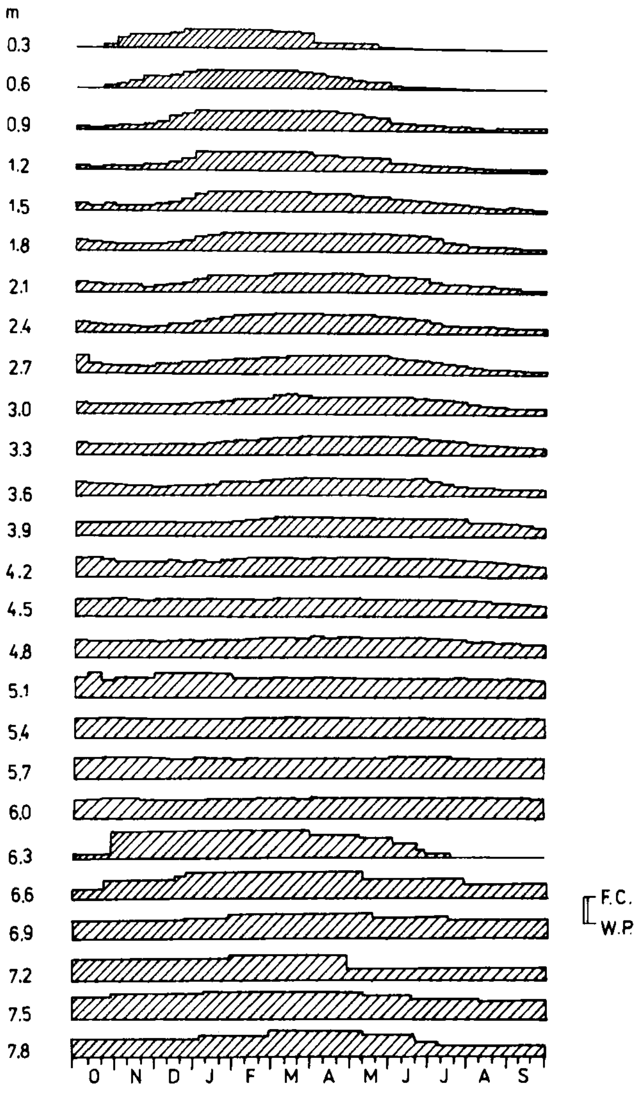
5.11 Representative basins in Uganda, Tanzania and Kenya, serving for the assessment of hydrological régimes of the Great African Lakes.

changes in order to obtain a qualitative and quantitative evaluation of such changes on the hydrological régime. Of course, changes resulting from a spontaneous settlement of mankind, such as steadily rising agricultural and industrial production, urbanization etc., can be considered as contributing to the representativeness of the basin. Thus the only areas which are usually chosen as experimental basins are those subject to sudden changes in land use, vegetation and urbanization. Sometimes, however, the experimental areas serve for the testing of new measuring devices and methods, and for network design analysis. Also special features and phenomena can be measured and analysed in the experimental basins. Usually more than one experimental basin is established. Half of them are selected as experimental sites, while the other half serve as controls during experimentation. All basins used for experiments should be similar in terms of size, shape, topography, climate and other features. The natural behaviour of the basins should be observed for some years before any changes are made. The reference basin should not be subject to any induced or spontaneous changes during the experimental part of the project.



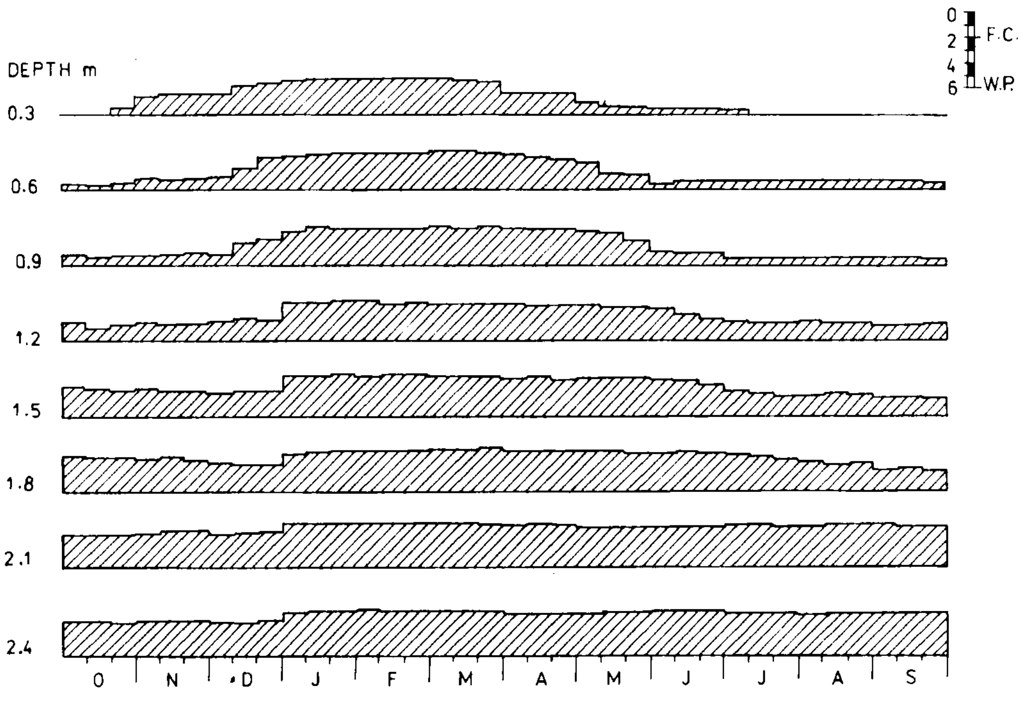
5.12 Small scale experimental catchments, Luano, Zambia. A dense network for observing the groundwater régime allows more definite conclusions to be drawn.

Fig. 5.12 gives an example of four experimental Luano catchments in northern Zambia. They were intensively observed for a period of three years, and then two of them were cleared of forest. The impact on the hydrological régime of turning forested land into farm land was studied in that region. The experiments demonstrated a high consumption of groundwater resources by tree roots when they reached the zone of capillary rise. After the clearing, the groundwater level rose high above the original mean value. Fig. 5.13 and Fig. 5.14 give a comparison of the soil moisture fluctuation under forest and under farm land.



5.13 Soil moisture fluctuations under tropical forest in wet and dry regions. Deep-rooted vegetation affects the soil moisture fluctuation down to the depth of 8 metres. Field capacity – F.C.; wilting point – W.P.

As a guide for the establishment and maintenance of representative and experimental basins, the book by Toebes and Ouryvaev [5] is very useful.



5.14 Soil moisture fluctuations under grassland in wet and dry regions. Shallow-rooted vegetation affects the soil moisture fluctuation down to the depth of 2 metres.

5.5 IDENTIFICATION OF CONSTRAINTS

Having collected all the existing information on the system, and the data available at present and expected in the future, the constraints which may affect the analytical part of the project have to be taken into account.

Most frequently, among the most typical constraints are lack of time, and limited budget and manpower. Limited instrumentation and computer facilities are often the follow-ups to these problems. One has to consider not only the constraints which appear in the initial stage of the project, but also the problems and limitations which may occur in the later stages of the work.

The time allocated for the project implementation needs to be defined very precisely. Hydrologists often depend on data and materials produced by other specialists, and therefore any deadlines for the completion of the hydrological part of the project should respect the condition that all necessary resources needed for the hydrological analysis will be available.

Often preliminary results are requested. However, meeting such requests may turn out to be a serious problem when, for various reasons, the final definitive results are different. For that reason, therefore, it is wise not to commit oneself to any preliminary conclusions.

When similar work has been carried out in the region before, the investor may attempt to compare previous results with the new ones, often ignoring the fact that new methods, measurements etc. can furnish entirely different results. The analysts, therefore, should acquaint them with all earlier works, and should be prepared to compare them with their own conclusions. Conservative investors are not always prepared for the application of advanced new methods, and may request a comparison between the new technologies and the old methods. Such a demand may lead to an increase in the volume of work originally contracted, and to misleading conclusions. This type of methodological constraint always should be foreseen at the initial stage of any project.

5.6 IDENTIFICATION OF WORKING METHODS

The choice of working methods depends upon the quality of data available, the amount of data which can be acquired within the framework of the project, and the feasibility of identifying the region, and it should respect all potential constraints. In some cases the preliminary simulation requested has to be simplified, and sometimes an application of the model has to be rejected when the available inputs do not meet the requirements of the model manual.

There are several general principles which are applicable under almost all circumstances. The most important principle is the unification of the period selected for data analysis. In some projects, the analysts try to make use of all data which has been collected and which are considered generally to be valuable material. However, only very problematic conclusions can be drawn from, for instance, the joint analysis of long rainfall records containing data from extremely dry and wet years, and much shorter discharge records available, say, for a single wet or dry period.

The length of the period selected for the analysis should be judged from two different points of view. The first is the length of observations records, available for most of the various observation points. At this stage all very short records, or records containing many gaps, will be eliminated. Remaining in the final selection will be all data from a period of observation which can be considered representative. In addition, a reasonably long record of several decades is needed for a few stations. It will serve for the assessment of long-term mean values, amplitudes, and the probability of occurrence of the extremes. While long records are usually available for rainfall observations, longer records of discharge, and particularly of groundwater level, are less common.

At least for the groundwater level records, the levels for the first and last day of the analysed period should be more or less equal, so that the impact of the groundwater storage deficit or excess can be precluded.

Provided that we intend to work with a three-year period of observation, a value close to the long-term mean can be obtained, for instance, from one very wet year and two dry ones or from one very dry year and two wet ones. Both periods, equally representative at first sight, may yield very different results. However, when we plan to establish a temporary observation network we can never be sure whether the proposed period will supply values comparable with the long-term means. Then the subsequent analytical work has to be concerned with the relation between the short-term period available, and the long-term records.

Concerning the groundwater régime, we have to evaluate the soil moisture régime. Apart from measuring the groundwater level fluctuation it is recommended to measure the soil moisture as well, so that the soil moisture deficit or surplus can be evaluated.

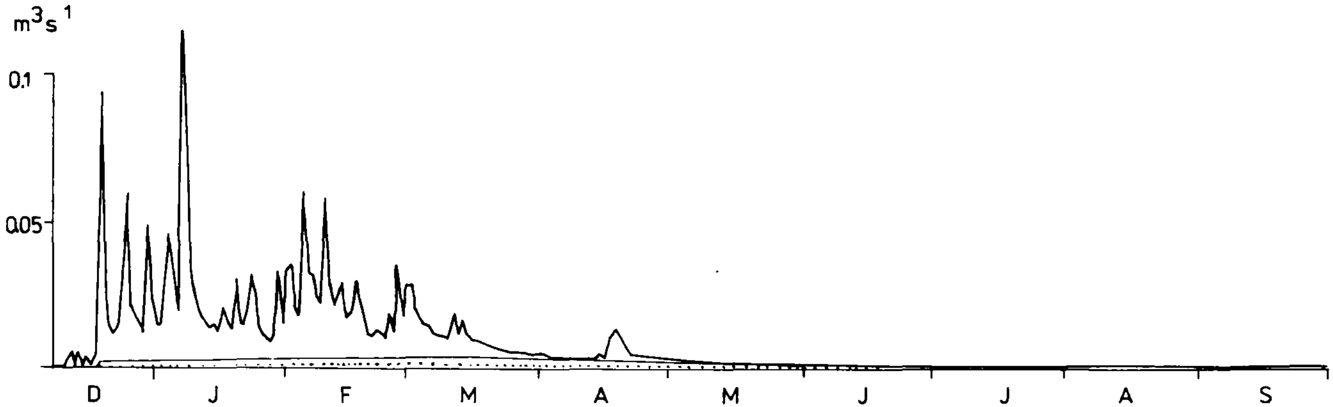
5.6.1 Hydrograph analysis

Determining the baseflow component is an important part of groundwater studies. When analysing a single hydrograph we are concerned with the results representing one area, whereas in the regional analysis we use the results to compare the behaviour of several subregions. Therefore the same method of analysis has to be used for all cross-sections observed in the region. In Fig. 5.15 is a comparison of two hydrographs and their components for tropical basins with intermittent régimes. In the first case 10.8% of the area consists of a shallow aquifer, while in the second case it is only 4.9%. Thus the surface runoff component which forms on the overfull shallow aquifer is considerably higher for the same precipitation régimes, and much less remains for the formation of groundwater outflow from the deep and shallow aquifers. Because in this case the vegetation is different for the ground surface of the deep and shallow aquifer, a conclusion can be reached on the rôle of land use in the formation of the baseflow.

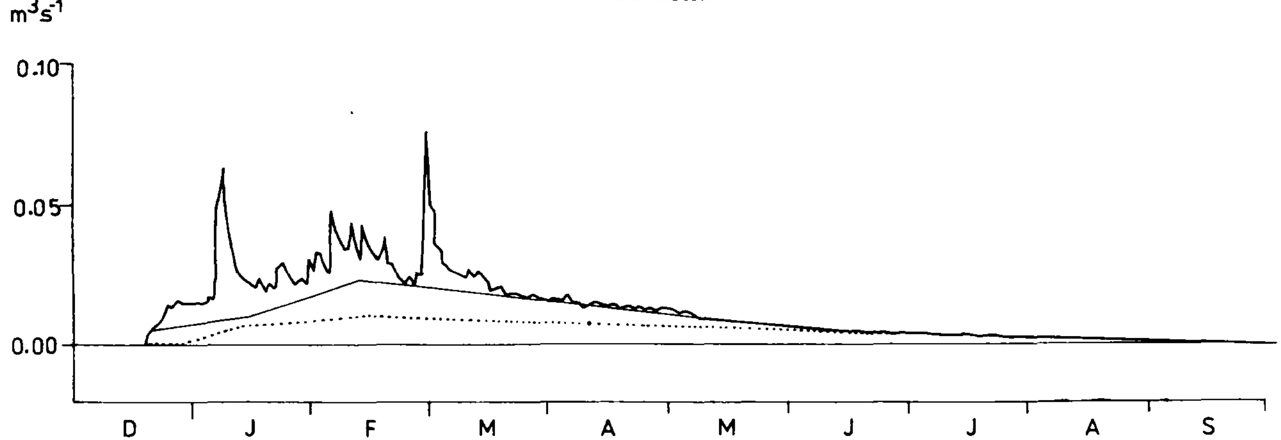
5.6.2 Rainfall-runoff relationships

The relationships between rainfall and runoff constitute an essential part of every analysis. If the baseflow component can be determined, then a modified relationship can be plotted for rainfall and baseflow. In Fig. 5.16 are plotted the rainfall-baseflow relationships for tropical catchments with higher and lower percentages of area exposed to the surface runoff. In the first case, the baseflow formation is less sensitive to rainfall than in the second case, where owing to reduced surface runoff the conditions for groundwater recharge become more favourable.

Such a conclusion is valid for a particular case only; we can arrive at different results in other regions. For instance, a steep gradient of the rainfall-runoff curve may indicate a fast recharge of the aquifer from the rainfall water, and it is possible

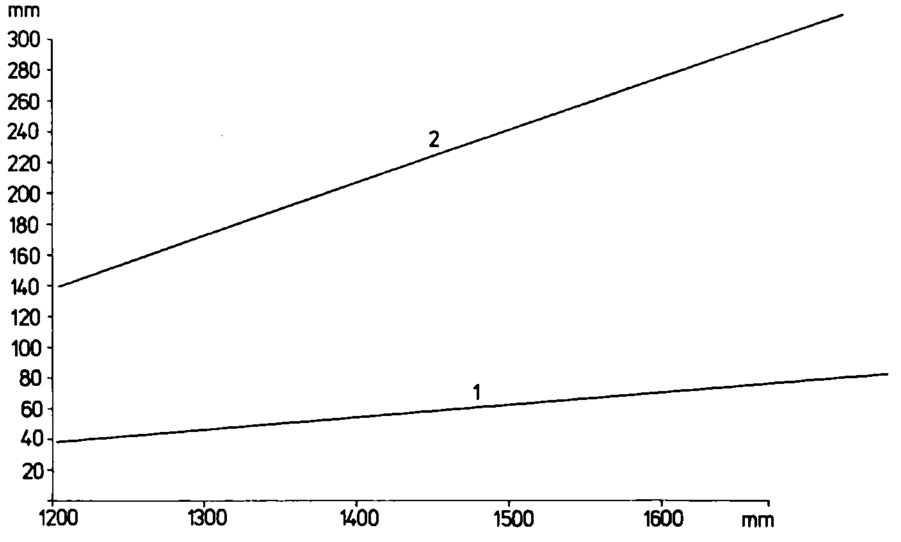


5.15.1 Hydrograph from the catchment containing 10.8% of shallow aquifer. The higher surface runoff component and reduced groundwater components (as compared with the hydrograph in Fig. 5.15.2) should be noted.



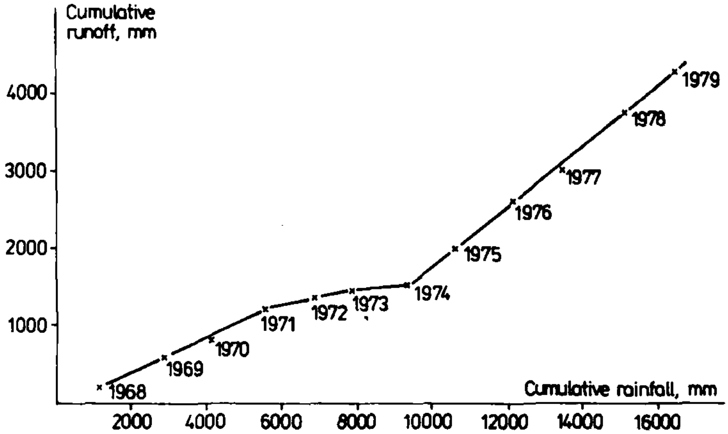
5.15.2 Hydrograph from the catchment containing 4.9% of shallow aquifer. Here the conditions are less favourable for the formation of surface runoff, and the groundwater component is better developed.

that the groundwater storage is highly dependable on the rainfall régime. Another extreme would be a line parallel with the rainfall axis, which would indicate that the groundwater resources were independent of the rainfall. Probably a great volume of groundwater is stored there, and the pondage effect of the subsurface reservoir is very pronounced. This does not necessarily mean that the aquifer could never be recharged by rainfall water. It only indicates that the annual recharge volume is relatively small compared with the volume of stored groundwater. A source of confined water can be manifested in a similar way.



5.16 Rainfall–baseflow relationships for the catchments with 10.8% of shallow aquifer – *1*, and for the catchment with 4.9% of shallow aquifer – *2*, based on the annual water balance.

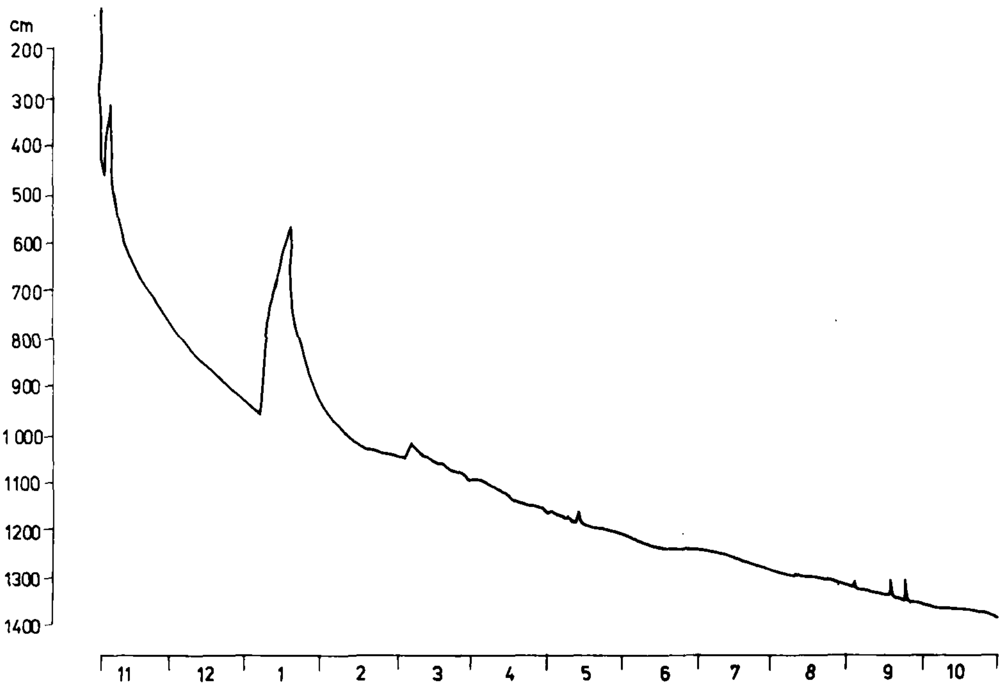
In some cases we may find it more convenient to apply what is called the double-mass curve. This curve can adjust the changes in the hydrological régime which are caused by changing factors or conditions. Instead of plotting annual rainfall and runoff totals, the cumulative totals for a period of several years are plotted. In Fig. 5.17 is the double-mass curve for the total rainfall and runoff. This graph was used to test whether the runoff régime of the Luano catchments in northern Zambia had been changed due to the clearance of the forest. In 1973/74 a sudden change in the slope of the line can be traced, indicating an immediate impact of forest clearance. Because the increased slope of the curve represents the increased surface runoff, we can conclude that after the forest had been cleared, much less of the rainfall water remained available for the baseflow.



5.17 Double mass curve of the total rainfall and runoff. The change in the régime after deforestation began in 1974.

5.6.3 Analysis of the groundwater régime

In the very initial stage, we are concerned with the régime which can be considered characteristic of a given area. When comparing two or more graphs of the ground-water-level fluctuation, very rarely do we find completely identical patterns. In



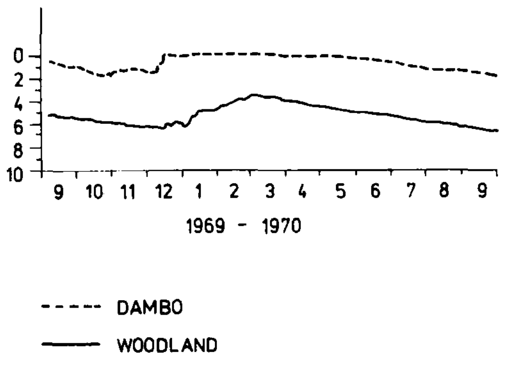
5.18 Groundwater-level decline after a one-year pumping test. An amount in excess of the estimated safe yield was extracted in the area. A temporary rise indicates an interruption in the pumping.

non-homogeneous aquifers, under semi-confined conditions, with a rolling or hilly surface and a bedrock gradient, a great variety of fluctuation patterns can be discovered. The first task is to find their common features which are distinct for a given type of aquifer as compared with another one.

Initially latent man-produced effects have to be identified, such as the prolonged effect of pumping, or recovery of the aquifer after pumping (Fig. 5.18). It is sometimes difficult to trace the effects of temporary pumping or small-scale pumping tests, particularly when the groundwater recharge is very sensitive to the precipitation régime.

Any gaps in the groundwater level records have to be filled in carefully, because usually these records are not long enough. It is always difficult to find out whether the mean level in a short analysed record is identical with the long-term mean. Is should be noted that, frequently, the mean representative rainfall period is not identical with the mean groundwater level period, particularly in regions with low rainfall totals. Here the groundwater level rises only during the so-called "wet years" with higher precipitation.

Amplitudes of the seasonal peaks which differ from borehole to borehole do not always indicate different régimes. The varying porosity of the zone of saturation, and variation in the gradient of the groundwater level on a regional scale, are among factors which may contribute to differences in the amplitude. Usually differences in the timing of rise and fall of groundwater level, as shown in Fig. 5.19, will indicate different types of groundwater régime.

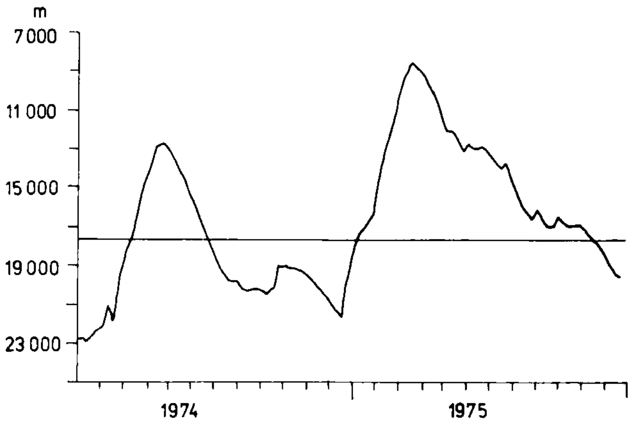


5.19 Two types of groundwater régime in the same region. The upper line is the groundwater-level fluctuation in a shallow aquifer with a fast response to the rainfall, while the lower one is typical for a deep aquifer under woodland with a slower response to the rainfall.

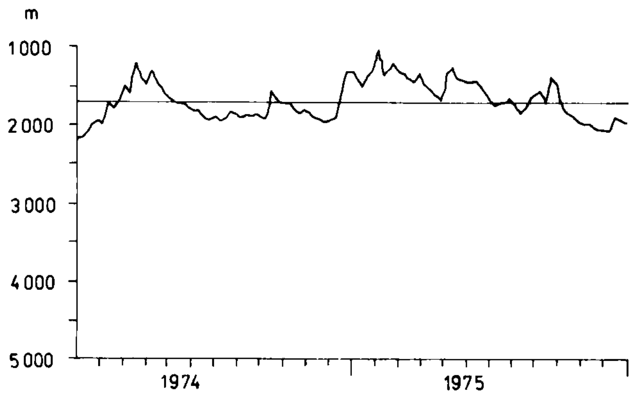
If the groundwater level begins to rise early during the beginning of the rainy season, as compared with a delayed start, it indicates in the first case that the zone of aeration is rather shallow, and therefore becomes saturated soon after the first rains arive. If the groundwater level is stable during a prolonged rainy season, it indicates that the aquifer is rather shallow and overfilled. To the contrary, the recharge of the aquifer with a steadily rising groundwater level indicates that the régime is not limited by the aquifer storage capacity.

The depletion of both aquifers will differ as well. Thus the position of boreholes with such different régimes can lead to the identification of subregions within a studied area.

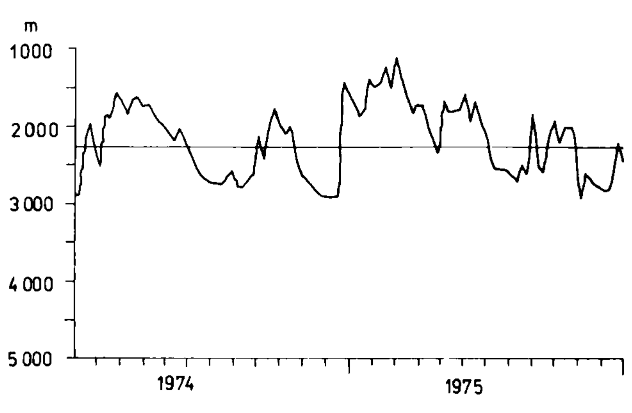
Fig. 5.20 contains a comparison of three different régimes in a temperate climate.



5.20.1 A comparison of the groundwater régime fluctuation in three aquifers of the same region: Deep aquifer with seasonal recharge pattern.



5.20.2 Deep aquifer under less developed soil profile and repeated groundwater recharge throughout the year.

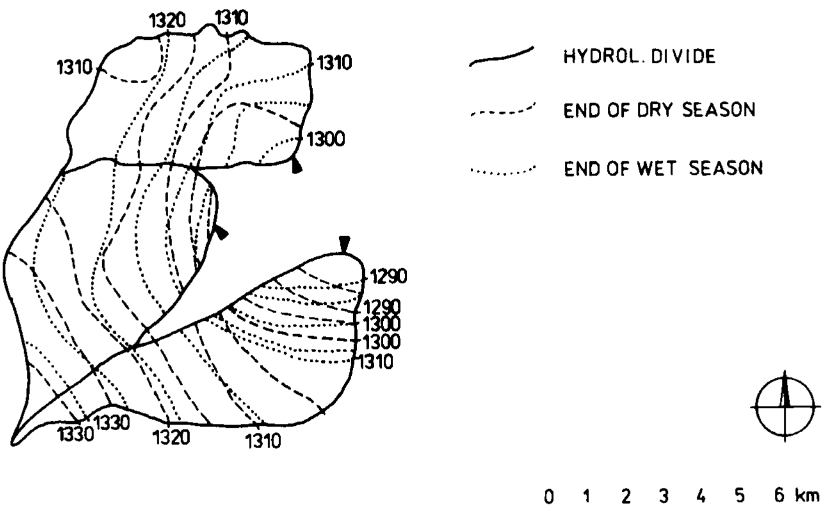


5.20.3 Shallow aquifer with frequent groundwater recharge after each heavy rainfall.

In the first case, the regular fluctuation of the groundwater level without any secondary peak indicates a rich live storage in a deep aquifer, with a well-developed soil profile and clear seasonal recharge. In the second case the groundwater recharge occurs more often, and secondary peaks indicate a fast response of the groundwater to precipitation. An all-year-round recharge after each rain and a fast depletion can be seen in the third case. Here, quite probably the contribution to the baseflow from such a shallow aquifer is limited to short periods during a year.

A stable groundwater level suggests that the groundwater is not interacting at all with the precipitation on the one hand, and the stream network on the other. Alternatively, of course, a continuous stable flow of groundwater may exist under the stream network. In such a case the groundwater storage is very probably replenished by exogenous sources. A comparison of the groundwater-level altitude in distant boreholes will provide additional information.

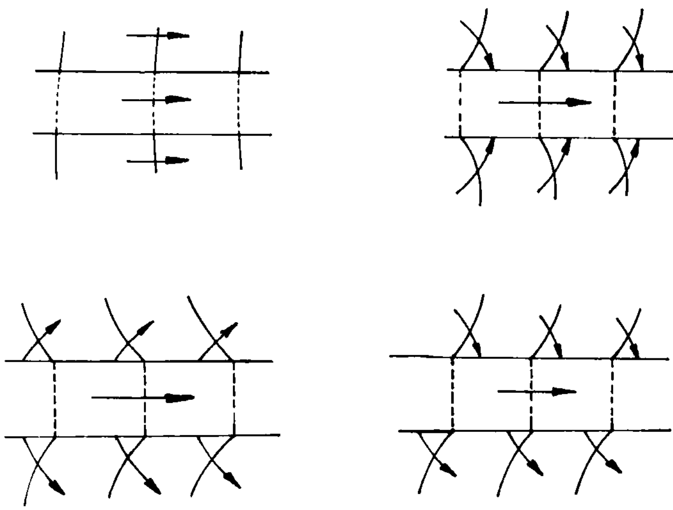
Contour maps of the water table or of piezometric surfaces and perpendicular flow lines, favoured in hydraulic models, can be successfully analysed only when a sufficient number of boreholes are observed; otherwise, from a small number



5.21 Contour charts of the water table in the Luano catchments, Zambia, at the end of the dry season and at the end of the rainy season.

of measuring points, misleading results will result. If such a contour map is based on data from boreholes with fluctuating groundwater levels, only measurements taken within the same time interval must be used. A comparison of the contours obtained for the typical seasons of a year can provide valuable information (Fig. 5.21).

During a flood period in the region, groundwater levels in the shallow aquifer are temporarily raised by inflow through the river banks and the stored water is called the bank storage. Sometimes it is difficult to establish whether the fluctuation of the water in a shallow well has occurred due to recharge from precipitation or as a seepage from the river network. Then a comparison between the groundwater-level régime in a well located near the stream, and that in a distant well, may yield useful information. Todd [6] recognized four different interactions between stream and groundwater (Fig. 5.22).



5.22 Four different interactions between groundwater and stream; after Todd.

Groundwater-level fluctuations can be caused by phenomena other than precipitation. In particular, water levels in boreholes which penetrate confined aquifers may be influenced by the atmospheric pressure, ocean and earth tides, earthquakes and construction works. Even if these impacts are infrequent, all possibilities have to be taken into account when analysing the groundwater régime.

5.6.4 Analysis of the régime of springs

Fluctuations in the régime of springs often provide valuable supporting information; sometimes the springs may become a main object of interest. It is usually difficult to identify the drainage area of a spring and thus the yield has to be judged very carefully. From one point of view springs belong to the groundwater, from another they can be considered as part of the surface system.

A spring is a concentrated appearance of groundwater on the ground surface. Sometimes a spring is defined as a current of flowing water; however, many springs remain hidden because, for instance, the outlet is under the stream surface. As opposed to springs with their concentrated discharge, seepage areas are recognised as a separate phenomenon. The seepage water at the surface may pond, flow or evaporate. Somewhere between springs and seepage areas belong the so-called mound springs found in Australia. According to Williams and Holmes [7] the mound water is dissipated by a mean evaporative flux of about 5.5 mm day^{-1}. The term "mound" is related to a distinct pattern caused by the accretion of sediments around the outlets, the height of which may exceed 10 metres.

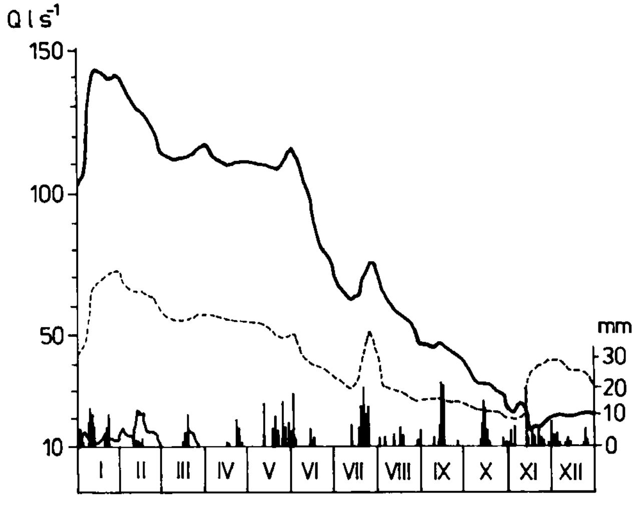
The classification of springs is concerned with the rock structure, discharge régime, water quality régime and temperature. In principle, springs resulting from nongravitational and from gravitational forces are distinguished.

Gravitational springs include: depression springs with the outflow at the intersection of the land surface and groundwater table; contact springs formed in a permeable aquifer overlying less permeable formations; artesian springs; and fracture springs which issue from fractures in impermeable rock strata, lava tubes, etc.

Nongravitational thermal springs are often associated with a fissure system extending to great depths. High mineralisation is a typical feature of thermal springs.

Perennial, intermittent and ephemeral springs can be recognised, in the same way as for streams.

Seasonally stable, perennial, springs are linked with extensive aquifers or are supplied by exogenous sources of water. Springs with a low discharge variability are normally found in volcanic rocks and sandstone formations. In mountainous areas, at high altitudes springs are found which exhibit fluctuations associated with the rainfall régime. An example of this is the annual fluctuation of the Sulkovy spring (Kříž [8]) in the hilly region of central Czechoslovakia (Fig. 5.23).



5.23 Fluctuation of the Sulkovy springs in Central Czechoslovakia, associated with the rainfall régime. After Kříž.

Periodic thermal springs resulting from the pressure of superheated steam at great depths, or from eruptions of gassy groundwater, are called geysers. They may indicate areas of marked tectonic activity.

When the chemical composition of springs differs from that of the groundwater system, it may serve as an indicator that the water originated in different strata.

In general, the spring régime can be analysed in a similar fashion to the régime of surface streams.

5.6.5 Analysis of man's activity in the region

Almost everywhere the surface régimes are more or less influenced by the activity of man. Not only organised activities but also spontaneous ones (resulting from the presence of the population and its growth), contribute significantly to changing the hydrological régime. Man's impact on groundwater resources is sometimes more difficult to establish, because of the time delay between the human activity and its impact on groundwater. In many cases the effect becomes irreversible when, for instance, the aquifer is overexploited for a considerable period of time. Besides the effects on water quality, which are beyond the scope of this book, major changes can be caused by an excessive withdrawal of large volumes of water. An example of such overexploitation through mining was discussed by Alföldi [9]. Irreversible effects caused by excessive pumping for irrigation, water supply, etc., have been reported from many parts of the world.

Any possible impact, past, present and future should be analysed in the initial stage of the project. The following types of activites which may possibly influence the groundwater resources are presented as a guide:

1. Spontaneous effects which result from a growing population,
2. Impacts resulting from changes in land use and agriculture,
3. Impacts resulting from industrial development,
4. Impacts resulting from increassing urbanisation.

According to Teller [10], any impact analysis must include a description of both the environment to be affected and the proposed actions on it. It is to be expected that the latter are more readily defined than the former which may require considerable inter-disciplinary study before it can be adequately described. An example of such approach was given by Chuanmao [22] who studied the groundwater régime fluctuation under the effects of human activities in arid and semi-arid parts of north China.

Because of the complexity of the problem the analytical part is discussed in Chapter 8. In the preliminary stage we have to be able to analyse and predict changing behaviour of the groundwater régime under the various types of human activities foreseen in the region.

5.7 IDENTIFICATION OF THE SUBREGIONS

Very rarely do we find a project area which behaves uniformly as a single hydrogeological unit. Therefore, the results of the previous analyses have to be used to identify those subregions within the project boundaries which, from the hydrological point of view, can be considered as behaving uniformly.

First of all, climatically uniform subregions have to be identified. Differences in climate can be expected only across large areas, so under normal circumstances the region can be considered as a single unit in terms of climate. The rainfall pattern analysis, particularly the seasonal distribution of the rainfall, will provide basic guidance.

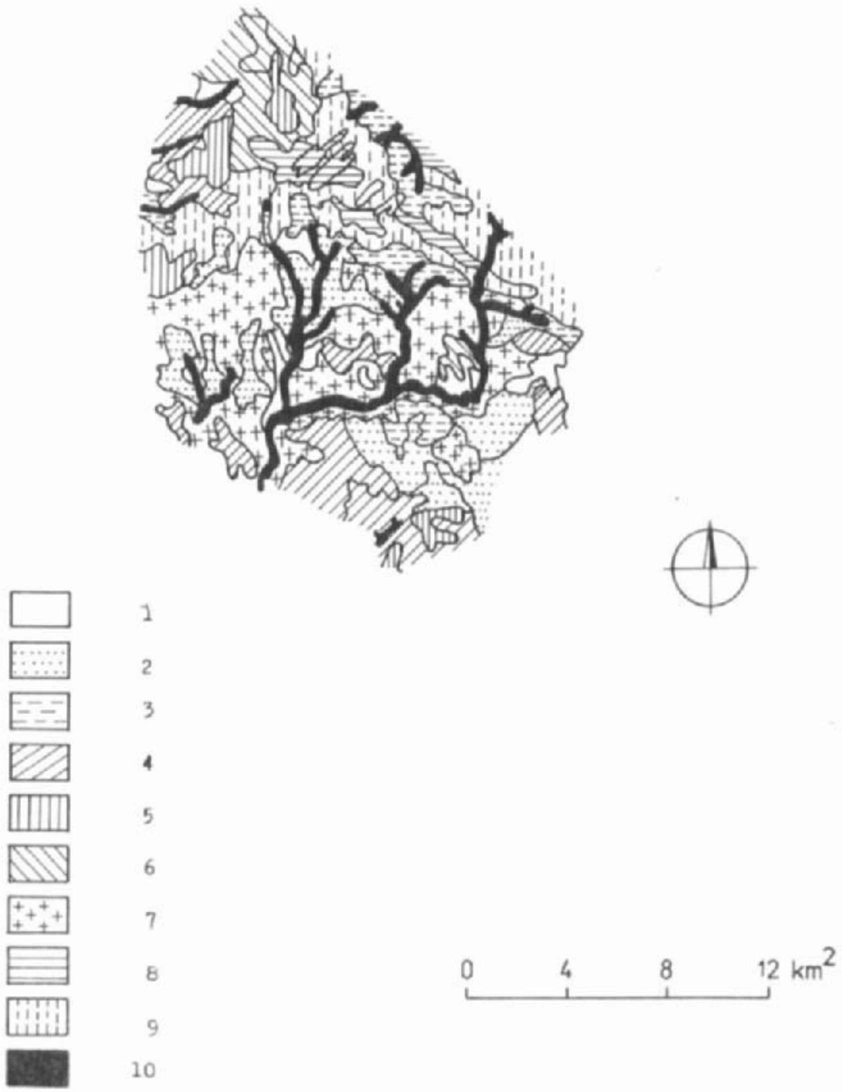
In the second step, areas with similar groundwater régimes have to be identified. For this purpose a comparison between the groundwater régimes in the observed boreholes is used, and in some cases the vertical stratification of the groundwater systems has to be taken into account.

In some regions a seemingly uniform groundwater régime can exist beneath a highly variable ground surface; in which case, when attempting to identify areas with uniform recharge régimes, preferred pathways and impervious areas, further investigation has to be concerned with the topography, soil and vegetation. Fig. 5.24 gives an example of the variety of soil which can overlie a Turonian aquifer. The recharge processes in each of the units are quite probably different from each other. Still, however, some common features can be found when establishing a large number of subregions would not be feasible.

The vegetational pattern plays an equally important rôle. Until recently, an areal impact of vegetation on the hydrological cycle has been one of the marginal fields of hydrology. The importance of vegetation has become widely recognised after the development of remote sensing methods and their introduction into hydrology. By identifying the vegetation and its changes in various bands, the processes of evapotranspiration and soil moisture fluctuation, and regionalisation aspects can be traced. Fransworth et al. [14] listed the possible applications of remote sensing identification in hydrogeology, particularly in the regionalisation approach. Remote sensing can be used to:

1. find indicators of the presence of groundwater,
2. indicate the quality of groundwater recharge and discharge,
3. indicate regions of groundwater recharge and discharge,
4. suggest areas where wells might be drilled, and,
5. monitor aquifer changes as groundwater development proceeds.

Some of the surface indicators of groundwater which can be recognised by remote sensing are: favourable geological structures, evidence of past surface water activity, the existence of transpiring vegetation in unirrigated areas, and the presence of active discharge from seepages and springs.



5.24 A variety of soil types overlying a Turonian aquifer in Central Czechoslovakia. Light deep brown soil – *1*; illimerised soil – *2*; brown acid soil – *3*; deep illimerised soil – *4*; podzolised soil – *5*; forest podzolised soil – *6*; acid soil – *7*; deep brown soil – *8*; shallow brown soil – *9*; alluvial deposits *10*.

Remote sensing has been successfully employed for regionalisation and groundwater exploration in Australia (Ellyet and Pratt [11]) using visible and near-infrared bands. Analysis of the hydrological régimes in arid regions was carried out by El Shazly et al. [12]. Guglielminetti et al. [13] identified coastal springs in the coastal

Tab. 5.4 Water cadastre of the Upper Congo basin

River	Cross-section	Drainage area	Rainfall	Runoff	Loss	Runoff coef.	Yield	Q	Discharge likely to be exceeded for ...% of a year				
		km^2	mm	mm	mm		$\frac{l}{s\,km^2}$	$\frac{m^3}{s}$	20	40	60	80	95
Chambeshi	above Wiwa Kalungu	5 986.0	1138	254.0	884,0	0.223	8.1	48.2	108	52	24	6.1	1.2
Wiwa Kalungu	to mouth of Chozi	3 911.5	1016	137.2	878.0	0.135	4.0	17.0	33	9.0	4.0	1.1	0.2
Chozi	mouth to Wiwa Kal.	5 493.8	1046	195.6	850.4	0.187	6.2	34.0	76	36	14	4.2	0.8
Wiwa Kalungu	below Chozi	9 395.3	1031	172.0	859.0	0.167	5.4	51.0	109	45	18	5.3	1.0
Interbasin Chambeshi	m. of Chozi-Chambeshi below Wiwa Kalungu	576.5	1041	121.9	919.1	0.157	3.9	2.2	4.8	1.6	0.8	0.3	0.1
Wiwa Kalungu	mounth to Chambeshi	9 971.8	1034	168.4	856.6	0.163	5.3	53.2	114	47	19	5.6	1.1
Chambeshi	below Wiwa Kalungu	15 957.8	1069	200.2	868.9	0.187	6.3	101.4	222	99	43	12	2.3

waters of Italy. Attempts were also made to identify alluvial deposits with huge groundwater reservoirs where confining beds have been formed.

Indicators of the existence and quality of groundwater near the surface are provided by vegetation growing near discharge points, or having roots which extract water from an aquifer. Where seeps, springs or seasonal lakes exist, the residue from evaporated water will show evidence of minerals.

5.8 PRELIMINARY WATER BALANCE ANALYSIS

The water balance calculation is, at a later stage, one of the final results of the groundwater resources assessment. In order to obtain a preliminary hydrological picture of the region studied, it is useful to calculate basic and simplified water balance components for such significant points as confluences, gauged cross-sections, lowest points of the basin etc. By this method the mean annual rainfall, mean annual runoff, and the "loss" components are calculated, together with the mean annual discharge and mean annual yield. If it can be estimated, the duration curves in the observed cross-section will provide valuable information. The probability of occurrence of floods is less frequent in groundwater studies; however, when groundwater and surface water are to be used together, flood analysis becomes highly valuable.

An example of the regional water balance table is given for the Upper Congo headwaters in Tab. 5.4.

By using a simplified water cadastre, other data can be obtained. For instance, the yield development, not only relevant to the mean annual discharge but also to the low values of the duration curves, indicates the development of the baseflow in the area.

5.9 SELECTION OR REJECTION OF THE MODEL APPROACH

Many projects these days are solved with the help of models in which the analytical solutions and their combinations are algorithmized. In fact, the main reason for applying the model approach is to simplify the analytical work, through a variety of alternatives based on variable inputs of all types. Therefore the choice of a suitable model should correspond with the information available. Sometimes, however, we can see project models used as a kind of decoration instead of serving as a useful tool for complex analytical work.

Numerous models have been constructed and are today available in all spheres of hydrology. Mathematical modelling in groundwater management at present covers a broad range of techniques, from algorithmized simple analytical formulae to advanced numerical schemes. Each of these models has its place in groundwater management. New model approaches are being examined all over the world and new models have been set up. The amount of data available for any particular solution is rapidly increasing, and this, together with our improved ability to identify rather complicated systems of groundwater flow, facilitates the setting-up of increasingly sophisticated models.

The selection of the model most suitable for a given problem is not an easy task. The great complexity of problems, various types of groundwater exploitation, and the side effects of human activities and environment, are often found to be highly specific to the problem to be solved. Thus the potential model user frequently decides to develop a new model instead of searching among the models already tested.

There are several possible ways of classifying groundwater models (Balek [15]).

5.9.1 Classification based on the type of a modelled process

1. General models simulate the groundwater régime as part of the hydrological cycle, and as a result of the continuous exchange of water within the atmosphere-plant-soil-groundwater-surface water system. Many general water balance models have been developed for various purposes, and those in which (a) the groundwater régime has not been considered a merely second or third priority process, and (b) is included the time of groundwater residence, can be considered useful tools.

The degree of complexity can vary enormously, and inputs varying from simple to sophisticated are involved.

2. Specialised groundwater models have been set up with the advent of the analog approach in simulation. Later digital versions have been developed, using methods of finite differences and finite elements. Some versions have turned out to be mathematically very complicated, far ahead of the inputs available.

3. Economic models simulate the strategy of groundwater resources exploitation in conjunction with water use. By using such models, environmental and water supply objectives can be analysed and optimized. Because of different levels of economic development in various regions, some formulae serve as guides rather than for direct application from one country to another. In some models, the land use impact is simulated as relevant to groundwater resources protection (Hardt, Hutchinson [16]).

5.9.2 Classification based on the mathematical approach

Basically four types of model can be classified in this group:

1. Black-box models solve the relationship between a given input and output when the physical background of the process is not fully understood. The transient functions stem mainly from practical experience of the behaviour of simulated variables. Greenwood and Burns [17] mathematically examined black-box and complex models, and concluded that the flow theory models require detailed data which are unlikely to be widely available and that, "contrary to many tacitly held beliefs,

models developed from the flow theory are not based on any more fundamental principles than the less complicated ones". For instance, a simple algorithmized relationship between groundwater level and baseflow can be considered a black-box model.

2. Conceptual models describe the physical processes in the catchment/field by empirical formulae supported by field and laboratory experiments. Each important physical process which takes place within the system is mathematically defined; the description usually reflecting recent knowledge of the existing processes and their interaction. An example of such a model is given in Chapter 6.

3. In contrast to conceptual models, statistical and stochastic models simulate the chance of the occurrence of events. Thus the concept of probability is introduced into the simulation. A model of a process which introduces the probability concept into the simulation is usually regarded as a statistical model. A model of the time series which takes into account the time factor as well as the probability, is considered a stochastic model. Several examples of these models are given in Chapter 7.

4. Dynamic models describe the simulated groundwater processes by a system of differential equations based on the flow theory. Sometimes two other categories are recognised, namely optimization models using linear and dynamic programming techniques, and dispersion models, which simulate the mass balance in situations of great hydraulic complexity. To the second group also belong the finite differences and finite elements models. As an example of the finite difference approach Trescott's model can be mentioned (Trescott et al. [18]). One of the most advanced finite element models has been proposed by Gupta et al. [19].

5.9.3 Classification based on the purpose of the models

These models are usually classified into three categories:

1. Models to promote scientific understanding are very important in the field of basic research, because they are effective tools for the promotion of a deeper knowledge of complex systems and also of specific problems within a complex system. Often, however, the amount and quality of the required inputs can be ensured only by experiments conducted in laboratories and in the field. With the increasing quality of field data some scientific models can be adapted for practical purposes.

It should be stressed that scientific models are not necessarily oriented toward the dynamic approach alone, and even less advanced mathematical solutions may lead to adequate scientific results.

2. Managerial models usually require some adaption of the scientific models for further utilisation. Here particular attention has to be devoted to the manuals which should be made available in a form which can be easily read by the potential user. The data and parameters involved also should be clearly defined, with an explanation of how to obtain them in the field or from literature.

3. Forecasting models can be divided into two types. One type provides prediction on a long-term basis and on a regional to global scale; the second type is concerned with short-term intervals and rather small areas. Both models exist at scientific and/or managerial levels, and any type of mathematical approach can also be applied. Sometimes long-term forecasting is based on three types of hypothetical data: optimistic, normal and pessimistic. Short-term forecasting is often performed in real time; in groundwater applications the intervals can be much longer than one day.

One of the dynamic forecasting models has been cited earlier (Hardt, Hutchinson [16]). Stochastic forecasting models are discussed in Chapter 7.

5.10 MODEL TESTING AND ADAPTATION FOR A REGIONAL PROBLEM

A quick appraisal of the many existing models is a difficult part of the identification approach, even for an experienced hydrologist. It has been stated by Prickett [20] that "... despite the impressive array of mathematical models and computational devices there are definite areas of needed improvements which include making present and future models useful by adequately documenting codes and procedures for application, developing a large group of models aimed at solving problems in the range of simple to moderate complexity, developing models and techniques that produce results a manager can understand". This statement is equally valid for any new models generated in the course of any project.

A Clearing-House for Groundwater Models which has been established at the Holcomb Research Institute, Butler University, Indianapolis, U.S.A., can provide initial advice for model selection.

If a model has been found convenient for solving regional problems, it may require further adaption for specific applications. Typically, this work may consist of re-organisation of the data bank, adaptation of headings and redistribution of the outputs and change of units.

It has often been found convenient to test the model in a control area located in the studied region. For this area the data are prepared with great accuracy. In this way the validity of the model can be tested, and it is important to do this even if the model builders claim that the model will not need any calibration. Also the credibility of the manual has to be verified, in relation to the computer system available.

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6 CONCEPTUAL ANALYSIS

6.1 GENERAL

In conceptual (deterministic) simulation, the hydrological cycle or parts of it are described by mathematical and/or empirical formulae. Because most of the formulae applied contain parameters characterising the physical features and properties of a modelled region, the conceptual models are sometimes called parametric models. In general it has been expected that a higher accuracy in the process description leads to more accurate models. From a practical point of view for a more complex description of a system more accurate inputs are required, and the quality of these may not come up to the expected level of the model. Therefore the complexity of the simulation should correspond to our ability to identify the system.

Model-builders sometimes differentiate between linear and non-linear models according to the type of formulae and equations used in the simulation. In another approach models with lumped and distributed parameters are recognized, and other types of classification are employed. Classifications according to type should be considered as secondary, and no model work should be initiated with the aim of setting up a system which will fit into a preselected type of classification. Provided we have a solid knowledge of the seasonal distribution of certain parameters and the ability to formulate a relevant formula, the latter should be introduced in the model, but otherwise a simpler, seasonally lumped parameter model is preferable.

If any part of the hydrological cycle is to be simulated by a conceptual model, the other parts of the cycle should not be omitted. The water cycle is dynamic and well balanced and all parts of the cycle are interrelated. This is, however, often left out of consideration by the model-makers. For instance, when a flood régime is conceptually solved for real-time forecasting purposes, far less attention is devoted to the interaction of soil, vegetation and atmosphere. Thus the significant effect of the vegetation functioning as a natural storage reservoir of intercepted water, of infiltration fluctuation, etc., is neglected, and a misleading concept will be inevitably introduced.

In a hydrogeological approach, the final output of a deterministic model would probably be an assessment of the safe yield as a product of water balance development between precipitation, baseflow, subsurface flow, overland flow and soil moisture, and groundwater excess and/or deficit. By using a complex conceptual approach we can obtain additional information on the size and function of groundwater reservoirs, interaction of groundwater and surface water, and the process of groundwater depletion and replenishment.

Numerous deterministic models simulating the water cycle have been developed in all fields of hydrology, but not many place an emphasis on groundwater resources, because in this field dynamic hydraulic models predominate. They simulate, with a high accuracy, the fluctuation of the groundwater level, which is often based upon approximations of the evaporation, recharge and boundary conditions. And so highly accurate mathematical methods are applied to very uncertain inputs.

Like other sciences, hydrology is in a state of rapid progress and advancement. A model which has been developed today can be found to be obsolete tomorrow. Therefore a conceptual model should be designed in order to allow a rapid exchange of the formulae used in the flow chart. Otherwise a great deal of additional effort may be spent on readapting the system. The same may be said about the data. We should bear in mind that models which are often tested by using data from specially equipped basins, have to be applied in regions where the data are scarce in quantity and poor in quality. It is recommended to set up a model which can work with approximate basic data when better input is not available, but an open channel for all additional data and information which can be obtained from other areas should be included in the simulation process.

Regarding hydrogeological application, we often hear the opinion that the water-balance approach is not fully applicable because of the special conditions prevailing in the groundwater system, such as long-term retention of the groundwater in the aquifer, long-distance transport of the groundwater to the stream network, and confined or semi-confined conditions. This is not true, because we can always simulate a groundwater system as one or several reservoirs of water, of which the mutual interactions can be described just like any surface hydrological phenomena. The geological conditions only determine the reservoir capacity, the type of water release, and water replenishment. Thus precipitation at the outcrop of a confined aquifer contributes to the volume of water already accommodated there, regardless of the time delay after which the water will be released into the stream network.

The first experiments with deterministic simulation were made in the early sixties. In 1963 Crawford and Linsley [1] published an article about their first deterministic model, in 1965 Dawdy and O'Donnell analysed a conceptual model approach [2], in 1967 a water balance conceptual model was described by Ayers and Balek [3, 4]. Since then a great number of various types of deterministic models have been constructed.

Because the principles of hydrological conceptual models are very similar, those who intend to apply a conceptual approach to solve hydrogeological problems may become confused by the variety of such conceptual models available, It is often difficult to choose whether to apply a model already developed, or to set up a new model which might fit a given problem better.

In the writer's experience, a certain type of model can always serve as a guidance when preparing a new approach. The parts of the model can be adapted and made more convenient for a given purpose, and familiar formulae can be applied for the solution of marginal problems. First of all, however, the quality and quantity of data available must be considered in the initial stage. In many models, the data input is arranged according to the situation typical for the model calibration, and this may be rather inconvenient for a given purpose. In a conceptual approach, the data input arrangement is almost always very significant, so it is strongly recommended to adapt at least that part of the model which concerns the data input, so that the data which are actually available can be fully utilised without any further adjustment.

To provide some guidance in the deterministic approach, an extended Guelph water balance conceptual model is described in this Chapter. It should be noted, however, that this description cannot fully replace the manual, and serves only as guidance for a conceptual approach.

The principles of the model were set up as early as 1967 (Ayers, Balek [5]) when the model was tested at the University of Guelph. Since then, however, many formulae have been changed, some of them several times, in accoradnce with the results of field experiments and model application results (Balek [6]). The original name was also changed to the Dambo Model, the native name for the African headwater swamps to whose régime the model had been applied. In what is known as the Wageningen version, the model was adapted to simulate the intermittent régime. Later another substantial modification was set up in order to assess the safe yield in extensive geological structures (Balek [7]).

6.2 MODEL CONCEPTION

The model was designed to meet the following criteria:

1. the ability to incorporate data from a great number of catchments;
2. to have a wide application; and,
3. compatibility.

For this purpose we require parameters characterising catchments, initial water balance values, and input and control data available for a preselected time interval. The model is adjusted for each of the time units chosen as a balance interval. Hourly intervals have been used as a basic time unit, but longer intervals can be applied upon request.

Although wide application is usually possible, the results depend on the quality of control data; as their accuracy decreases only approximate solutions can be obtained.

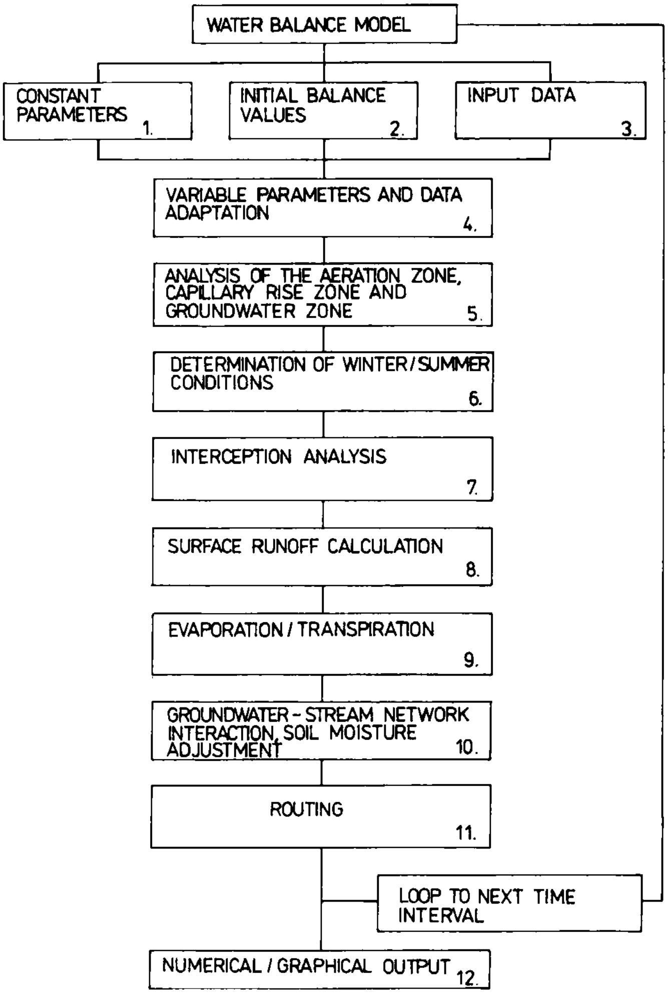
To achieve compatibility, the application of subroutines should be avoided. While the model itself is widely applicable, the subroutines concerned with the data organisation have to be set up according to the computer available. At this point some difficulties may be expected.

The following results are expected from the model:

1. a general picture of the hydrological cycle and its components;
2. a continuous water balance assessment of a general catchment over selected time intervals; and,
3. the testing of various methods and formulae proposed as characterising various parts of the hydrological cycle.

During the initial stage of modelling, when the model was set up as a tool for the study of the hydrological processes, the main objective was considered to be a general picture of the hydrological cycle in a given basin.

Water balance outputs are more pragmatic. They should be directly applicable for any practical purpose, for instance in agriculture and forestry. They can also be used as an input for water resources system models.



6.1 Simplified flow chart of the deterministic model.

The testing of various methods and formulae proposed as characterising water cycle processes is linked with both basic and applied research. A reasonable verification of formulae and methods can be provided only after the basic model has been proved to work reasonably well. The difference between the control and simulated data indicates the efficiency of new methods.

6.3 SYSTEM ANALYSIS

In general, any phase of the hydrological cycle can be simulated by the model for hourly and/or daily intervals. The results are presented numerically and/or graphically. The model is set up as a system which allows the fast exchange of formulae describing any part of the cycle. Provided that the balanced catchment is non-uniform ecologically, agriculturally and hydrogeologically, it can be divided into several subcatchments, each of them balanced separately.

The model is intended to simulate seasonal, spatial and other variations of the parameters characterising the catchment and/or parts of it. The artificial effects of pumping, recharge etc. can be simulated as well.

Real-time hydrograph forecasting, and real time forecasting of any balanced component, can be performed by the model if predictions of rainfall and air temperature are available. The model works as a simulation of the interaction between: the atmospheric situation, biological zone of interception, zone of surface runoff and snow accumulation, upper soil moisture zone, zone of capillary rise, groundwater zone, and the channel network, and it works under all summer and winter conditions.

A simplified flow chart of the model is shown in Fig. 6.1 and FORTRAN symbols for the variables and parameters are listed in Table 6.1.

Table 6.1 List of the symbols in FORTRAN for the conceptual model

General notes:

1. (*I*) is a symbol indicating number of the balanced subregion.
2. *J* indicates balanced hour.
3. *1*, *2* are the symbols for the values at the beginning or end of the balance interval, respectively.
4. Values given in % should be in such a form that 100% is the equivalent of 1.

Symbols:

A(*I*)	Actual value of the interception rate, reflecting also the conditions in the vegetation	mm $hour^{-1}$
AA(*I*)	Drainage area of the balanced subregion within the balanced region	km^2
AF(*I*)	Coefficient of the equation for the calculation of transpiration rate	–
AIXX(*I*)	Minimum value of the interception rate in a year	mm $hour^{-1}$
AIYY(*I*)	Maximum value of the interception rate in a year	mm $hour^{-1}$

AI(I)	Maximum value of the interception rate in a year when no water is intercepted	mm hour^{-1}
B(I)	Calendar-number of the day in which maximum development of the vegetation is expected	–
BF(I)	Coefficient of the equation for the calculation of the transpirational rate	–
DEPF1(I)	Depth of the frozen soil.	mm
DGW1(I)	Amount of groundwater storage in soil from the first day of balancing	mm
DHF(I)	Accumulated degree-hours below 0°C	°C
DIFA(I) *DIFB(I)* *DIFC(I)*	Coefficients of the differential equation characterising the surface runoff	
DLMA2(I)	Calculated storage in the zone of aeration	mm
DLMB2(I)	Storage in the zone of capillary rise	mm
DSNO2(I)	Water storage in the snowpack	mm
DUM2(I)	Water storage in the zone of interception	mm
E(I)	Potential evaporation observed or estimated as an input	mm interval^{-1}
EE(I)	Calculated total evapotranspiration from snow, interception, soil and water uptake by roots	mm interval^{-1}
ETP(I)	Potential evapotranspiration observed or estimated as an input	mm interval^{-1}
EP(I)	Total evaporation from snow and/or interception	mm interval^{-1}
ETPP(I)	Total evaporation from snow and/or interception	mm interval^{-1}
ETTP(I)	Evaporation from soil	mm interval^{-1}
ETS(I)	Evapotranspiration from soil and through the vegetation	mm interval^{-1}
F(I)	Instant initial interception rate	mm h^{-1}
FFF(I)	Capillary rise rate from groundwater to capillary zone	mm hour^{-1}
FMX(I)	Water release from the interception zone to soil in form of drops, stemflow, etc.	mm hour^{-1}
FFX(I)	Soil infiltration rate	mm hour^{-1}
FPOR(I)	Volume of capillary pores between field capacity and wilting point	%
GR(I)	Total groundwater outflow in balanced interval	mm hour^{-1}
GW2(I)	Groundwater storage	mm
HC(I)	Calculated groundwater level below soil surface	mm
HO(I)	Observed groundwater level below soil surface	mm
IDEPF1(I)	Depth of frozen soil when balancing starts	mm
IDHF1(I)	Accumulated degree-hours when balancing starts	mm
IGW1(I)	Dynamic groundwater storage when balancing starts	mm
ILMA1(I)	Soil moisture storage above capillary rise zone when balancing starts	mm
ILMB1(I)	Soil moisture storage in the capillary rise zone when balancing starts	mm
ILMPW1(I)	Amount of frozen water in capillary pores when balancing starts	mm
ILMWW1(I)	Amount of frozen water in noncapillary pores when balancing starts	mm
ISL(I)	Capacity of the soil moisture zone above the groundwater level	mm
ISLA1(I)	Capacity of the zone of aeration when balancing starts	mm
ISR1(I)	Initial surface runoff	mm
IT(I)	Time delay in the start of the surface runoff hours	
IUM1(I)	Initial storage in the zone of interception	mm
IX	Calendar number of the first day of balancing	

K(I)	Recession coefficient	
KK(I)	Constant correcting the evapotranspiration accordingly to the active part of the subregion	
LMA1(I)	Soil moisture storage above capillary rise zone	mm
LMB1(I)	Soil moisture storage in the capillary rise zone	mm
LMPW1(I)	Amount of frozen water in capillary pores	mm
LMWW1(I)	Amount of frozen water in noncapillary pores	mm
M(I)	Water melted from snow in the water balance interval	mm $hour^{-1}$
MEANQ(I)	Mean daily groundwater outflow yield	$m^3 s^{-1} km^{-2}$
MEANQA(I)	Mean daily groundwater outflow	$m^3 s^{-1}$
MEANQO(I)	Mean daily total discharge – observed	$m^3 s^{-1}$
MEANQC(I)	Mean daily total discharge – calculated	$m^3 s^{-1}$
MEANQRA(I)	Mean daily discharge from surface runoff	$m^3 s^{-1}$
NAR	Number of subregions	
ND	Number of days in a year	
NPOR(I)	Volume of noncapillary pores	%
P(I,J)	Precipitation	mm
PI(I)	Artificial recharge/depletion	$m^3 s^{-1}$
PW(I)	Volume of capillary pores available for frozen water	mm
Q(I)	Groundwater outflow yield	$m^3 s^{-1} km^{-2}$
QA(I)	Groundwater outflow	$m^3 s^{-1}$
QC(J,I)	Total calculated discharge	$m^3 s^{-1}$
QO(J,I)	Total observed discharge	$m^3 s^{-1}$
QR(I)	Surface runoff yield	$m^3 s^{-1} km^{-2}$
QRA(I)	Surface runoff	$m^3 s^{-1}$
R	Soil resistence against freezing/coefficient	
RH(I)	Instant depth of roots	mm
RHXX(I)	Minimum depth of roots in a year	mm
RHYY(I)	Maximum depth of roots in a year	mm
S	Number of seconds in balance interval	
SDEE(I)	Total evapotranspiration from the first day of balancing	mm
SDGR(I)	Amount of groundwater outflow from the first day of balancing	mm
SDP(I)	Precipitation total from the first day of balancing	mm
SDSR2(I)	Total surface runoff from the first day of balancing	mm
SG(I)	Groundwater zone capacity-active part	mm
SH(I)	Accumulated standard error of estimate for *SHO* and *SHC*	
SHC(I)	Mean daily groundwater level below soil surface, calculated	mm
SHO(I)	Mean daily groundwater level below soil surface, observed	mm
SI	Insulation capacity of snow cover in mm of water content	mm
SL(I)	Instant capacity of capillary pores above groundwater level	mm
SLA1(I)	Instant capacity of capillary pores above the zone of capillary rise	mm
SLB(I)	Capillary zone capacity	mm
SNO1(I)	Water content of the snowpack	mm
SQ(I)	Accumulated standard error of estimate for *QO* and *QC*	mm
SR1(I)	Surface runoff volume	mm
SU(I)	Instant capacity of the intercepted zone	mm
SUMA(I)	Result of the balance equation used as a control	mm
SUME(I)	Potential evaporation from the first day of balancing	mm
SUMEP(I)	Evaporation from surface runoff, snow, and interception from the first day of balancing	mm

SUMETP(I)	Potential evapotranspiration from the first day of balancing	mm
SUMETS(I)	Evaporation from soil and vegetation from the first day of balancing	mm
SUMETTP(I)	Soil evaporation from the first day of balancing	mm
SUMP(I)	Precipitation total from the first day of balancing	mm
SUMSR(I)	Surface runoff total from the first day of balancing	mm
SWW(I)	Amount of capillary water accumulated or depleted from the groundwater storage from the first day of balancing	mm
T(I)	Groundwater outflow time of travel	hours
TR(I)	Surface runoff time of travel	hours
TT(I)	Air temperature	°C
UM1(I)	Intercepted water storage	mm
WPOR(I)	Volume of capillary pores at the wilting point	%
WW1(I)	Capillary water in the groundwater zone depleted or accumulated due to the groundwater level fluctuation in balance interval	mm
XX(I)	Minimum capacity of the interception zone in a year	mm
X	Calendar number of the balanced day	
YY(I)	Maximum capacity of the interception zone in a year	mm

By changing the parameters which characterise the physical properties of the catchment, one can also simulate the impact of environmental changes on the catchment régime. There follows a list of the balanced values and instantaneous values of the variable parameters supplied in hourly intervals (Tab. 6.2):

Tab. 6.2 List of data supplied at hourly intervals

Water balance data

1. Water storage intercepted – *UM2(I)*, mm,
2. Surface runoff volume – *SR2(I)*, mm,
3. Soil moisture in upper moisture zone – *LMA2(I)*, mm,
4. Soil moisture in the zone of capillary rise – *LMB2(I)*, mm,
5. Groundwater storage – *GW2(I)*, mm,
6. Snow-pack water equivalent – *SNO2(I)*, mm,
7. Water storage in soil noncapillary pores – *LMWW2(I)*, mm,
8. Water storage in soil capillary pores – *LMPW2(I)*, mm,
9. Groundwater outflow *QA(I)*, $m^3 sec^{-1}$, and yield – *Q(I)*, $m^3 sec^{-1} km^{-2}$,
10. Surface runoff *QRA(I)* and yield – *QR(I)*, $m^3 sec^{-1}$,
11. Total evaporation from snow and/or interception – *EP(I)*, mm,
12. Total evapotranspiration from snow, interception, soil and water uptake by roots – *EE*, mm,
13. Potential evaporation/evapotranspiration – *E(I)*, *EPT(I)* (input), mm,
14. Groundwater level observed and simulated – *HO(I)* and *HC(I)*, mm,
15. Discharge observed and simulated – *QO* and *OC* $m^3 sec^{-1}$.

Variable parameters

1. Instant interception rate – *A(I)*, mm $hour^{-1}$,
2. Instant interception capacity – *SV(I)*, mm,

Tab. 6.1 cont./2

3. Instant depth of the roots – *RM(I)*, mm,
4. Initial infiltration rate – *F(I)*, mm hour^{-1},
5. Degree-hours – *DHF(I)*, °C
6. Depth of soil freezing – *DEPF2(I)*, mm.

For daily intervals, values in Tab. 6.3 are calculed as the differences between initial and instant days of balancing.

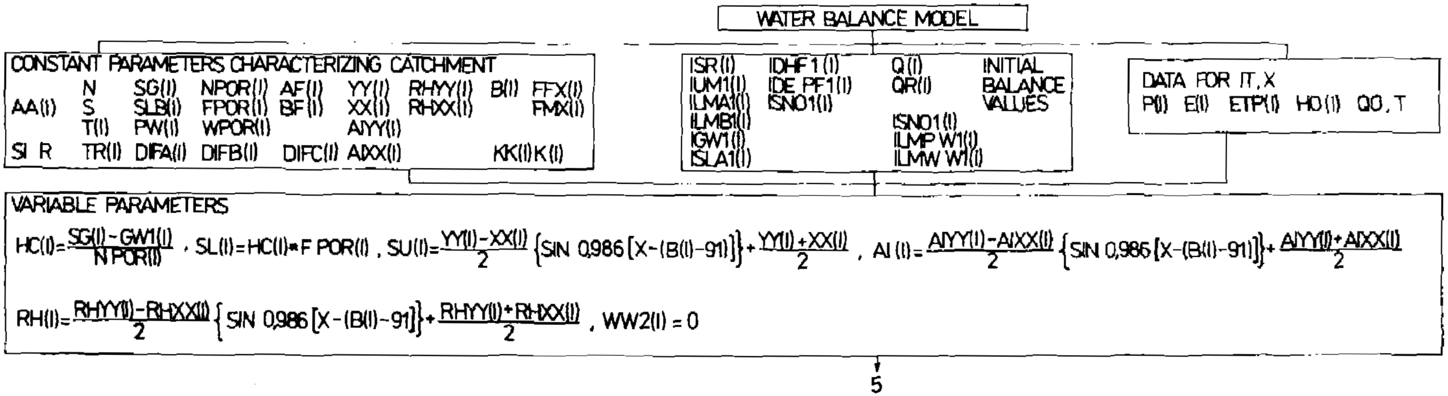
Tab. 6.3 Daily balance values supplied by the model

1. Total precipitation – *SUMP(I)*, mm,
2. Total potential evaporation – *SUME(I)*, mm,
3. Total potential evapotranspiration – *SUMETP(I)*, mm,
4. Total calculated evaporation from interception and snow cover – *SUMEP(I)*, mm,
5. Total evaporation from soil *SUMETTP(I)*, mm,
6. Total evapotranspiration from soil and through vegetation – *SUMETS(I)*, mm,
7. Total evapotranspiration, mm, including interception and snow – *SDEE(I)*, mm,
8. Total surface runoff – *SDSR2(I)*, mm,
9. Total groundwater outflow – *SDGR(I)*, mm,
10. Storage change in the zone of interception – *DUM2(I)*, mm,
11. Storage change in the upper moisture zone – *DLMA2(I)*, mm,
12. Storage change in the zone of capillary rise – *DLMB2(I)*, mm,
13. Storage change in the groundwater zone – *DGW2(I)*, mm,
14. Storage change in the snow-pack water content – *DSNO2(I)*, mm,
15. Storage change in the capillary pores of groundwater zone – *SWW(I)*, mm,
16. Control balance values which should be equal to total precipitation *SDP(I)* = *SUMP(I)*; when system works properly – *SUMA(I)*, mm,
17. Mean daily groundwater level observed – *SHO(I)*, mm,
18. Mean daily groundwater level calculated – *SMC(I)*, mm,
19. Mean daily groundwater yield – *MEANQ(I)*, m^3 sec^{-1} km^{-2},
20. Mean daily groundwater outflow – *MEANQA(I)*, m^3 sec^{-1},
21. Mean daily surface runoff – *MEANQRA(I)*, m^3, sec^{-1},
22. Mean daily observed discharge – *MEANQO(I)*, m^3, sec^{-1},
23. Mean daily calculated discharge – *MEANQC(I)*, m^3 sec^{-1}.

6.4 MODEL INPUT

There are three sources of data serving as the model input:
1. parameters characterising catchments;
2. initial balance values; and,
3. hydrological and meteorological data.

Assuming that the simulated catchment is divided into NAR subcatchments which are ecologically, agriculturally and hydrogeologically uniform, and that input



6.2 Flow chart of inputs and variable parameters. Partial flow chart 1 – 4.

varies from 1 to NAR, input is to be provided separately for each subcatchment (I). The maximum number of subcatchments in the present version is 10. The partial flow chart for the model input is shown in Fig. 6.2.

It should be noted that the initial values of the parameters usually enter the model as lumped, however, provided another information on their distribution is available, formulae for distributed parameters can be easily added. The subregions $AA(I)$ represent our image of the region divided horizontally and/or vertically. Here the adaptability of the areal representation gives the analyst a tool for a flexible approach.

All porosities can be initially obtained from manuals, handbooks etc. However, some field measurements are recommended to adapt them to the real conditions which prevail in the studied region. The same applies for all parameters representing the vegetation. Data on the leaf index and its seasonal changes provide good initial information.

Initial groundwater storage $IGW(I)$ is computed for the first day of calculation using the formula

$$IGW(I) = \frac{QA(I)\,3.6}{-\ln K(I)\,AA(I)} \tag{6.1}$$

where $QA(I)$ is the observed/estimated baseflow from the subregion $AA(I)$ and $K(I)$ is the recession coefficient determined for that region by hydrograph analysis. Then the groundwater storage capacity $SG(I)$ can be determined for a given subregion and first interval:

$$SG(I) = HO(I)\,NPOR(I) + IGW1(I) \tag{6.2}$$

where $HO(I)$ is the observed mean groundwater level in the subregion, $NPOR(I)$ is the noncapillary porosity, and $IGW(I)$ is the value calculated before. In the absence of information on mean groundwater level in the subregion, data from a single borehole or several boreholes can be utilised.

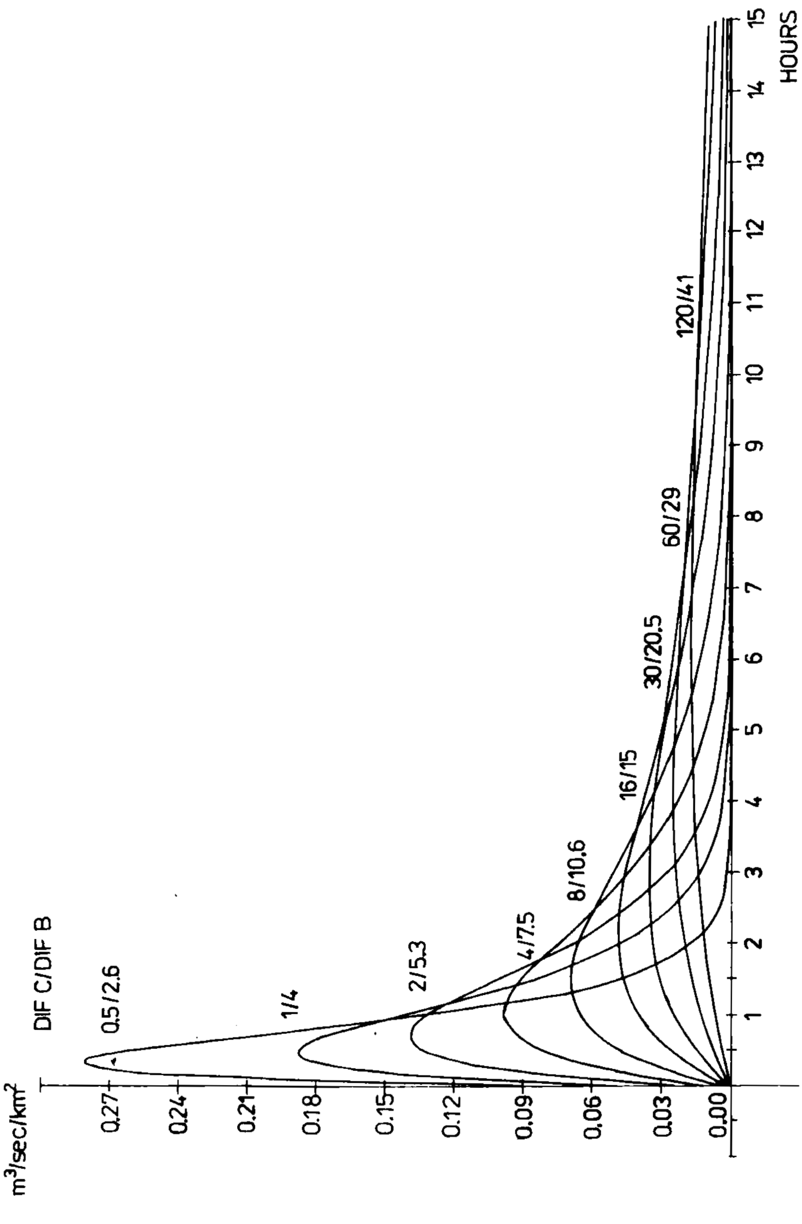
For the routing of the surface runoff the following differential equation is used (Ayers, Balek [4], Balek, Jokl [13]):

$$DIFC(I)\frac{\mathrm{d}^2[QR(I)]}{\mathrm{d}t^2} + DIFB(I)\frac{\mathrm{d}[QR(I)]}{\mathrm{d}t} + DIFA[QR(I)] = SR2(I) \tag{6.3}$$

In the applied metric system $DIFA(I) = 3.5$. The optimum ratio $DIFC(I)/DIFB(I)$ can be estimated with the help of Fig. 6.3.

All other input parameters can be found for the first run in the handbooks, in the reports of regionally oriented research projects, etc. Alternatively, if time permits, they can be provided by field measurements.

It will be found advantageous to start the water balance calculations when most of the water balance values for the first interval are equal to zero. The most suitable time for starting is the end of a hydrological year, when the soil is relatively dry



6.3 Optimum ratio DIFC I/DIFB I for solving the rainfall–runoff differential equation.

and air temperature still above zero. Because $IGW1(I)$ has been already calculated, only the capacity of the zone of aeration $ISLA1(I)$ needs to be determined, using the formula:

$$ISLA1(I) = ISL(I) - SLB(I) = HO(I)\,FPOR(I) - SLB(I) \qquad 6.4$$

where $SLB(I)$ is the capacity of the zone of capillary rise, $ISL(I)$ is the capacity of the soil moisture zone above the groundwater level, $HO(I)$ is observed groundwater level at the beginning of balancing, and $FPOR(I)$ is the capillary porosity. Initial soil moisture values $ILMA1(I)$ and $ILMB(I)$ should be based on the assessment of the soil moisture deficit for the first time interval. In long-term applications of the model, an initial deviation of the soil moisture from an measured value will not play any significant rôle.

Input data enter the model in hourly intervals; if they are available only in daily intervals, an arrangement should be made when preparing the databank. The same is valid for any other interval.

The size of the subregions should be such that the observation network will supply representative mean values, or at least data from single observation points. It may be sufficient to extrapolate meteorological data from neighbouring areas.

In order to find mean values, the database should be able to supply various types of alternative combinations in numerical and graphical forms. Therefore one of the most important, and sometimes the most difficult, parts of the conceptual approach is the organisation of the database which provides the input.

In most cases, only information on the potential evaporation $E(I)$ is available, while information on the potential evapotranspiration $ETP(I)$ is less commonly known. The figures for potential evaporation are then used as basic information instead of $ETP(I)$. Sometimes we may find it more feasible to store in the database the data on air temperature, air humidity, wind velocity, and radiation/sunshine, and calculate the evaporation/evatranspiration on line.

It should be noted that while precipitation, evaporation, evapotranspiration and air temperature are data which are used directly in the calculations, groundwater level and discharge are control data which serve only to calibrate the model and evaluate the results. In fact, the water balance simulation can be performed without them, but with lowered accuracy.

6.5 VARIABLE PARAMETERS

In the partial flow chart 4 (Fig. 6.2) the variable parameters are adapted for the balanced time interval. In particular, the vegetation should be adapted according to its seasonal development. Instant interception zone capacity $SU(I)$, interception rate $AI(I)$, and root development $RH(I)$ are calculated using the formula:

$$ISV = \left(\frac{\mathrm{Max}V - \mathrm{Min}V}{2}\right)\{\sin 0.988\,[x - (B(I) - 91)]\} + \left(\frac{\mathrm{Max}V - \mathrm{Min}V}{2}\right) \tag{6.5}$$

where ISV is the instant value for the balanced day, $\mathrm{Max}V$ and $\mathrm{Min}V$ are maximum and minimum values in a year, and $B(I)$ is the calendar number of the day on which the development of ISV is at its maximum (Ayers, Balek [5]). Many other formulae can be followed, based on the leaf area index (Knissel [8]).

Here also groundwater level $HC(I)$ is calculated for each time interval:

$$HC(I) = \frac{SG(I) - GW1(I)}{NPOR(I)} \tag{6.6}$$

The expression is based on the final value of groundwater storage $GW1(I)$ from the preceding interval: $SG(I)$ is the groundwater storage capacity and $NPOR(I)$ is the noncapillary porosity.

Provided that, instead of using spatially lumped parameters of any other type we intend to use distributed ones, the calculation should appear in this partial flow chart.

It should be pointed out that the problem of lumped and distributed parameters often arises when conceptual simulation is applied. In general, if the information leading to spatial and/or time distribution is limited, lumped parameters should be introduced at first, as another approach may lead to an oversophisticated system which might be unable to supply adequate results due to the lack of realistic inputs.

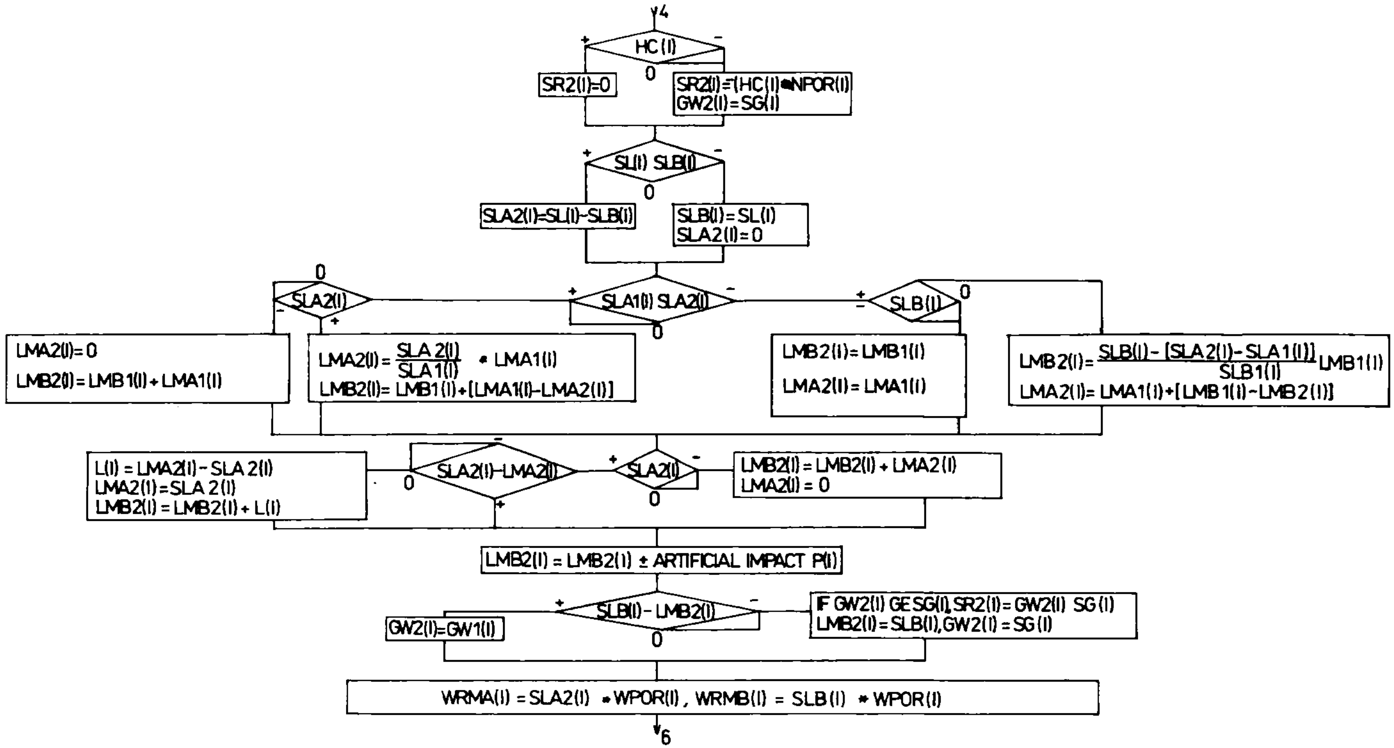
6.6 SOIL MOISTURE ZONES

The soil moisture fluctuations in the zone of aeration and capillary rise are simulated in the model, as can be seen in Fig. 6.4. They are far from being stable, and their size varies mainly in accordance with the fluctuation of the groundwater table. It should be noted, however, that the storage capacity of the zones represents mean values which correspond to the mean values of the groundwater level in the sub-region. As can be observed under natural conditions, the actual soil storage capacities from point to point can be found to differ greatly from those modelled (See subsection 7.8).

First a decision is made as to whether the groundwater level calculated in the preceding interval is still under the soil surface. If not, the excess groundwater will form the surface runoff "from the bottom" and thus the formation of swampy areas can be simulated.

Secondly, it is necessary to discover whether the groundwater level has been rising or falling. The zone of capillary rise fluctuates accordingly, and therefore the soil moisture in both soil zones will have to be redistributed. In the given version of the model the zone of capillary rise is assumed to be constant; again a distributed value can be employed, for instance, by introducing pF curves.

In the given version of the model, the possibility is also considered of human impact, such as the artificial infiltration or depletion of the water in the zone of capillary rise. This impact can be simulated in any zone. It is to be expected that the groundwater zone will be involved most frequently. It should only be noted that the impact

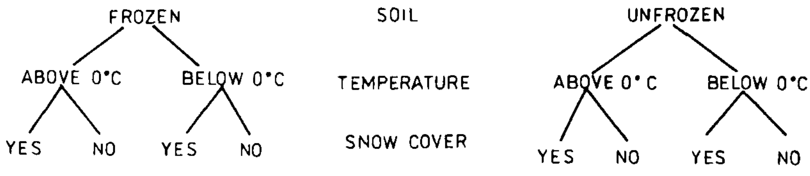


6.4 Soil moisture fluctuation in the zone of aeration and in the zone of capillary rise. Partial flow chart 5.

variable must be entered in the model in $m^3 s^{-1}$. The variable is stored in the database as a discharge, either positive or negative ($\pm PI(I)$).

6.7 WINTER CONDITIONS

Many conceptual models have been developed for summer conditions only. Particularly in the middle latitudes, the hydrological cycle is influenced by winter conditions for several months each year, and a reasonable conceptual model applicable in temperate regions should take this fact into account. It is the air temperature, soil temperature, and snowpack, which most influence the hydrological régime. A simple scheme in Fig. 6.5 indicates that in principle we may expect eight different hydrological régimes, depending on whether the soil is frozen or not, the



6.5 Types of winter/summer régimes.

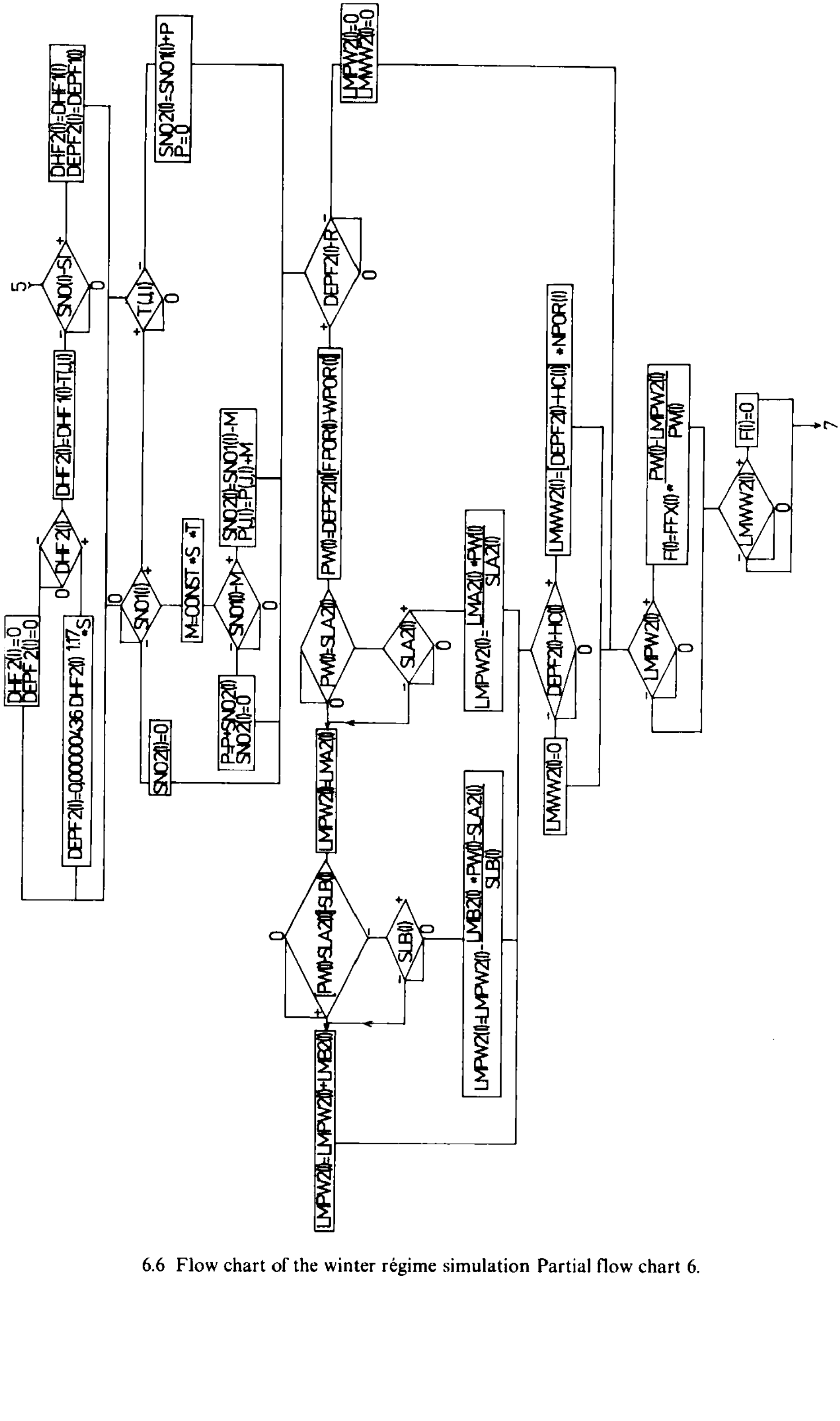
temperature below or above 0° C, and the subregion covered by snow. In fact, only one of these régimes can be considered as a purely summer régime, and all the others are more or less winter régimes. There is no doubt that various types of winter conditions have a great effect on the formation of groundwater recharge, especially during the snowmelt when the soil is thawing.

Fig. 6.6 depicts a conceptual approach to calculating one type of winter/summer régime. Here the snow content is compared against the constant which characterises the snowpack insulation ability. If the snow cover is not sufficiently thick the degree-hours value is calculated and the depth of the soil freezing is recalculated:

$$DEPF2(I) = 0.000\,004\,36 DHF(I)^{1.17} S \qquad 6.7$$

where $DEPF2(I)$ is the depth of soil profile having a temperature below 0° C, in mm; $DHF(I)$ is the degree-hours value; and S is the number of seconds in the balance interval. This empirical formula was developed experimentally and can be improved for various soil types and latitudes (Balek [9]). For medium latitudes and rolling land the formula gives only approximate values; for degree-hours close to zero further improvement of the formula is foreseen.

When the air temperature is below 0° all precipitation contributes to the accumu-



6.6 Flow chart of the winter régime simulation Partial flow chart 6.

lated snow; for temperatures above 0° C the snowmelt rate is calculated by using another empirical formula:

$$M = CST(J, I) \qquad 6.8$$

where M is the snowmelt rate, in mm s^{-1}; C is a constant, at present equal to 0.000 008; S is the number of seconds in the balanced interval; and $T(J, I)$ is the air temperature. Then the actual precipitation enriched by the snowmelt is calculated.

Provided the soil is frozen below a critical value R, the redistribution of the frozen soil water in the zone of aeration and capillary rise can be calculated, depending on the depth of freezing of the soil profile. The flow chart is self-explanatory.

If the depth of the soil freezing reaches below the groundwater level, the amount of water frozen in noncapillary pores is calculated. For both events, the infiltration rate has to be recalculated and the conditions of groundwater recharge determined in accordance with the given type of winter/summer régime (Fig. 6.6).

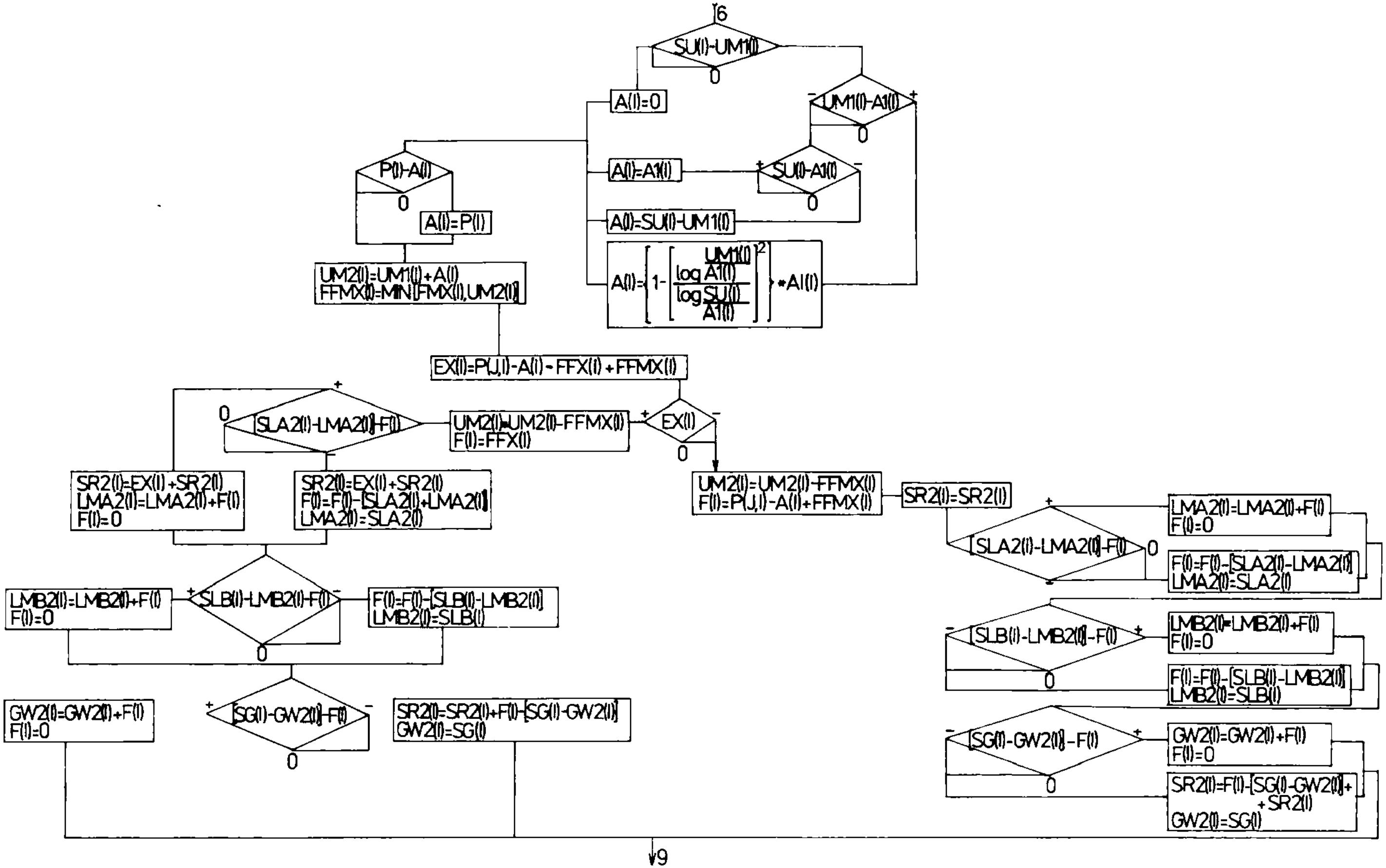
6.8 INTERCEPTION

Contrary to the opinion of some hydrogeologists who believe that the geological structure of the region is the dominant feature influencing the groundwater storage and depletion, the vegetation and the soil-plant-atmosphere process play a very important rôle in the formation of groundwater resources and the hydrological cycle. This is true at least from that moment when the precipitation comes into initial contact with the surface of the region to that moment at which the subsurface water finds a way to percolate through the noncapillary pores.

In between, there are many paths through which the intercepted and infiltrated water can be released back into the atmosphere. Nevertheless, greater attention has been paid to interception in models devised specially for agricultural and forest hydrology. In Fig. 6.7 there is a flow chart of the interception process, based on the presumption that the seasonal impact of the interception, and the variability of the interception rate, depend on the amount of water already stored on the surface of the vegetation (Balek [9]).

For a fully saturated storage capacity of the vegetation surface, the interception rate is equal to zero; when there is no intercepted water, this rate is at its seasonal maximum. An empirical formula based upon experiments in Czechoslovakia (Balek, Holeček [10]) and in Canada (Ayers, Balek [5]) is used: for the calculation of the instant interception rate $A(I)$, mm h^{-1}:

$$\frac{\log\left[\dfrac{UMI(I)}{AI(I)}\right]}{\log\left[\dfrac{SU(I)}{AI(I)}\right]} AI(I) \qquad \text{mm h}^{-1} \qquad 6.9$$



6.7 Flow chart for the simulation of interception. Partial flow chart 7 and 8.

where $AI(I)$ is the seasonal maximum interception rate, in mm h^{-1}, $SU(I)$ is the seasonal storage capacity of the vegetation, in mm; $UM1(I)$ is the amount of water stored as an interception, in mm. Obviously the condition

$$SU(I) > UM1(I) > A(I) > 0 \qquad 6.10$$

must be satisfied, otherwise the interception rate is defined by special statements. Using the same flow chart, we can calculate the instant amount of water flowing from the zone of interception to the soil surface as a stem flow, in the form of droplets, etc.

6.9 SURFACE RUNOFF

Fig. 6.7 shows solution of the process of surface runoff formation. It is based on an equation valid for each water balance interval:

$$EX(I) = P(J, I) - A(I) - FF(I) + FFMY(I) \quad \text{mm} \qquad 6.11$$

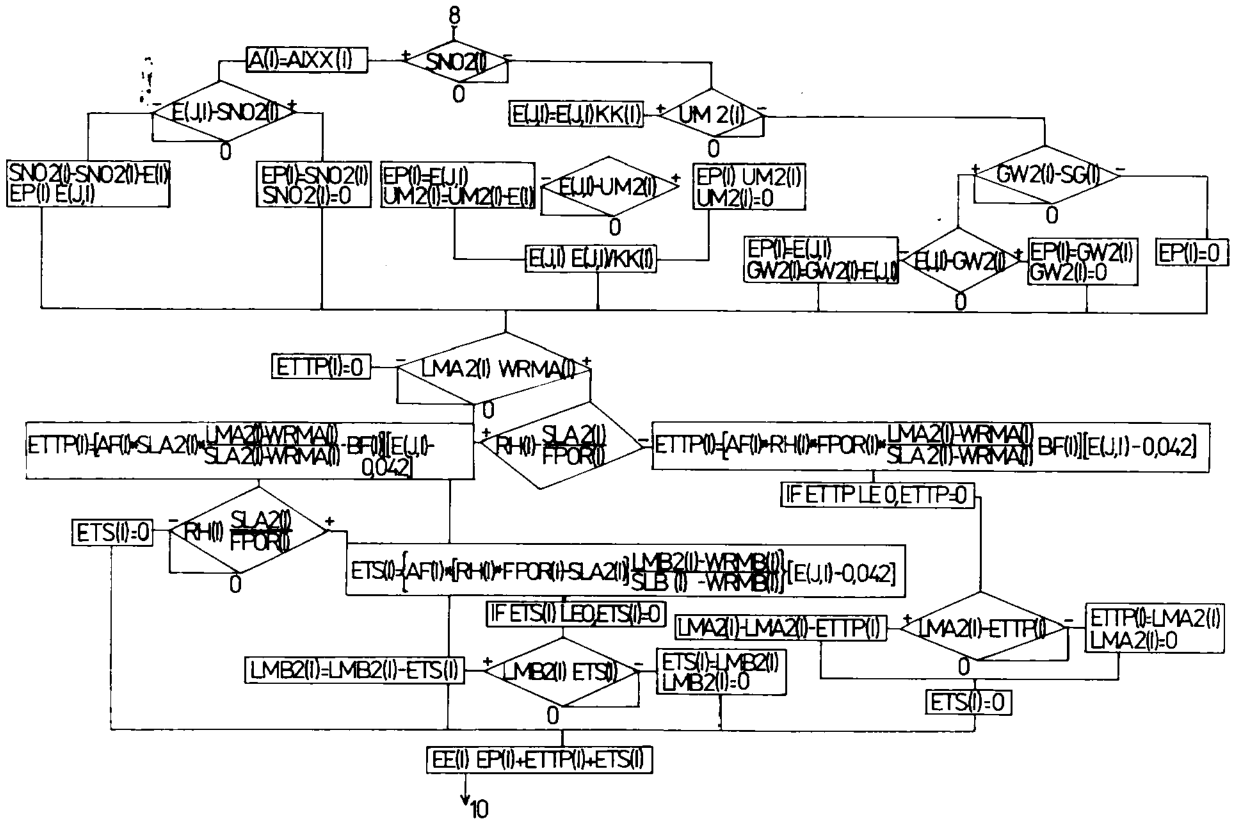
In this equation $EX(I)$ is the excess of water which remains on the soil surface, in mm; $P(J, I)$ precipitation, in mm; $FFX(I)$ is the soil infiltration rate, in mm h^{-1}; and $FFMX(I)$ is the rate of water transport from the zone of interception to the soil surface, in mm h^{-1}.

The soil infiltration is considered, for hourly intervals, as a constant because the most significant changes in the soil infiltration occur usually within the first 60 minutes. Provided that the excess $EX(I)$ is negative, the actual soil infiltration rate will be smaller than $FFX(I)$; then the infiltrating water can contribute first to the soil moisture in the zone of aeration and, when this becomes fully saturated, to the soil water in the zone of capillary rise and through it to the water storage in the groundwater zone. For extreme conditions the possibility of the formation of surface runoff is simulated, which, however, is more likely to occur when $EX(I)$ is positive. The surface runoff is then produced by the excess water on the soil surface, and in some cases additional surface water is contributed from an overstoraged groundwater zone. Infiltrating water can be fully or partly stored by the soil moisture zones, and fully or partly contributes to the groundwater storage.

6.10 EVAPOTRANSPIRATIONAL PROCESS

The flow chart in Fig. 6.8 has already been subject to several changes, and due to the rapid development of evapotranspiration formulae further alternatives are anticipated.

First the snow evaporation is solved, based on the potential evaporation for a given meteorological situation. If the surface is free of snow the evaporation of lakes and intercepted water, if any, is calculated, again using the potential evaporation.



6.8 Flow chart of the evapotranspirational process. Partial flow chart 9.

Here the impact of the vegetation development, which sometimes helps to reduce the evaporation and sometimes to increase it, can be considered using the coefficient $KK(I)$. The evaporation from soil is also based on the potential evaporation, provided water is available in the noncapillary pores.

It has been proved experimentally that transpiration may contribute to the total evapotranspiration, so that under certain circumstances the actual evapotranspiration is higher than the potential evaporation. This occurs especially under tropical conditions (Balek, Perry [11]). However, a temporary increase of transpiration was observed even in middle latitudes. An empirical formula used for computing the transpiration is based on the results of those experiments (Balek et al. [12]).

$$ETTP(I) = \left[AF(I)^* \, RH(I)^* \, FPOR(I)^* \frac{LMA2(I) - WRMA(I)}{SLA2(I) - WRMA(I)} \right] * [E(J, I) - 0.042] \tag{6.12}$$

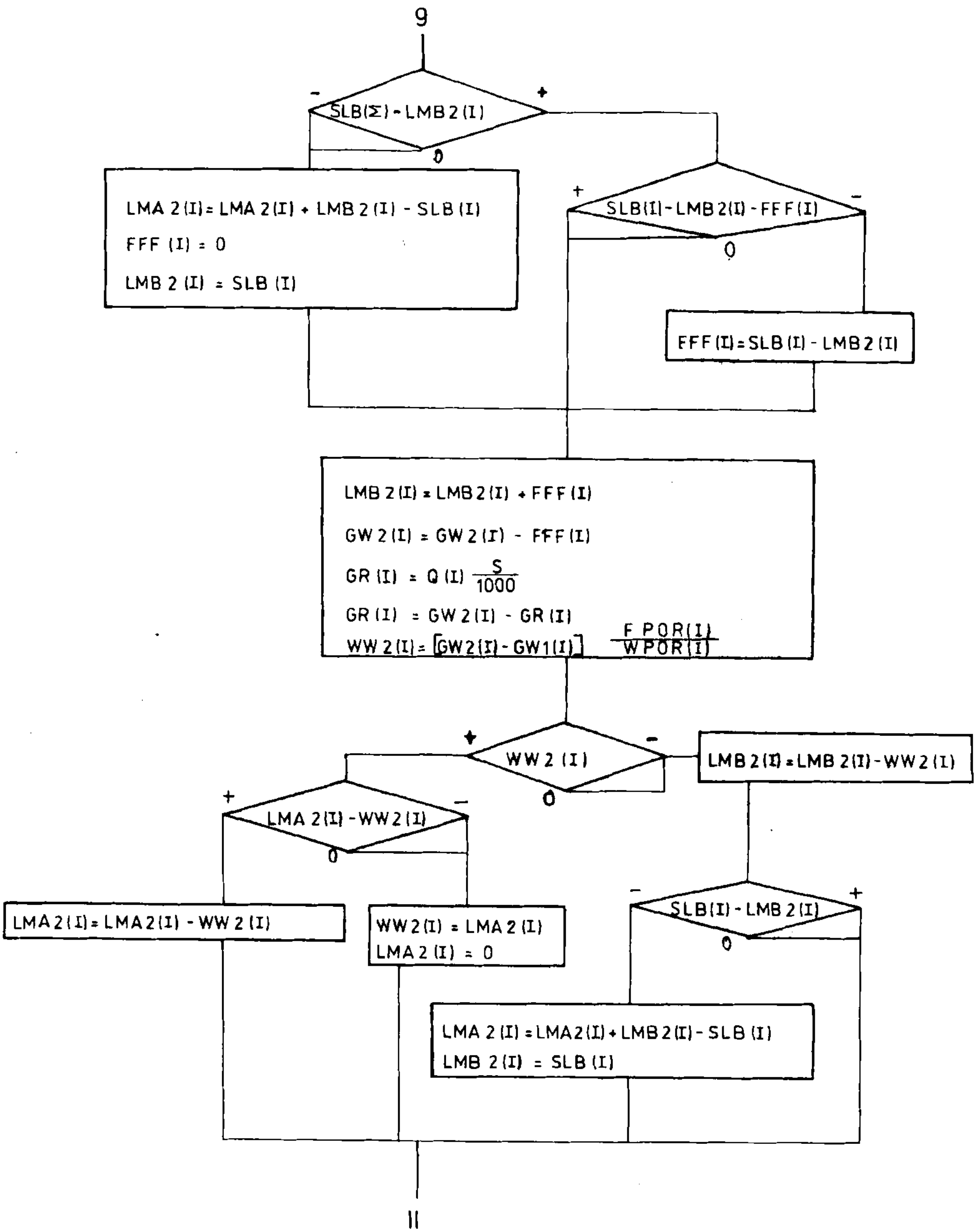
Here $ETTP(I)$ is the actual transpiration, in mm h^{-1}; $RH(I)$ is the seasonal depth of the roots, in mm; $FPOR(I)$ is the capillary porosity, in mm; $LMA2(I)$ is the amount of soil water in the zone of aeration, in mm; $E(J, I)$ is the potential evaporation for a given time interval, in mm; $AF(I)$ and $BF(I)$ are constants; and $WRM(I)$ is $SLA2(I) * WPOR(I)$, where $WPOR(I)$ is the porosity at the wilting point. When the root system extends to the zone of capillary rise, an analogical formula is used to assess the transpiration from that zone.

The total evapotranspiration is calculated as the sum of the evaporation from snow (water surface/interception) soil and the transpiration from both soil zones.

When potential evaporation appears in the transpirational formulae, it is used rather as an indicator of the meteorological situation than as an actual phenomenon. Even though experimental results always indicate a roughly linear relationship between transpiration and evaporation. this relationship varies from month to month and from year to year, probably according to the soil moisture content of the soil (Fig. 2.22).

6.11 GROUNDWATER–BASEFLOW INTERACTION, FLOW ROUTING AND MODEL OUTPUTS

As seen in Fig. 6.9, a continuous depletion of groundwater storage through an interaction with the stream network can be expected. The amount of water which has been released by the aquifer during the time interval is converted to millimetres ($GR(I)$) and subtracted from the groundwater storage calculated for the end of the interval. The final value of the groundwater storage $GW2(I)$ is compared with the initial groundwater storage $GW(I)$ at the beginning of water balance interval. Due to fluctuations of groundwater level, capillary water hidden in the groundwater



6.9 Flow chart of the groundwater-baseflow interaction. Partial flow chart 10.

zone $WW2(I)$ has to be added to or deducted from the soil water according to the movement of the groundwater level:

$$WW2(I) = [GW2(I) - GW1(I)]\frac{FPOR(I)}{NPOR(I)} \tag{6.13}$$

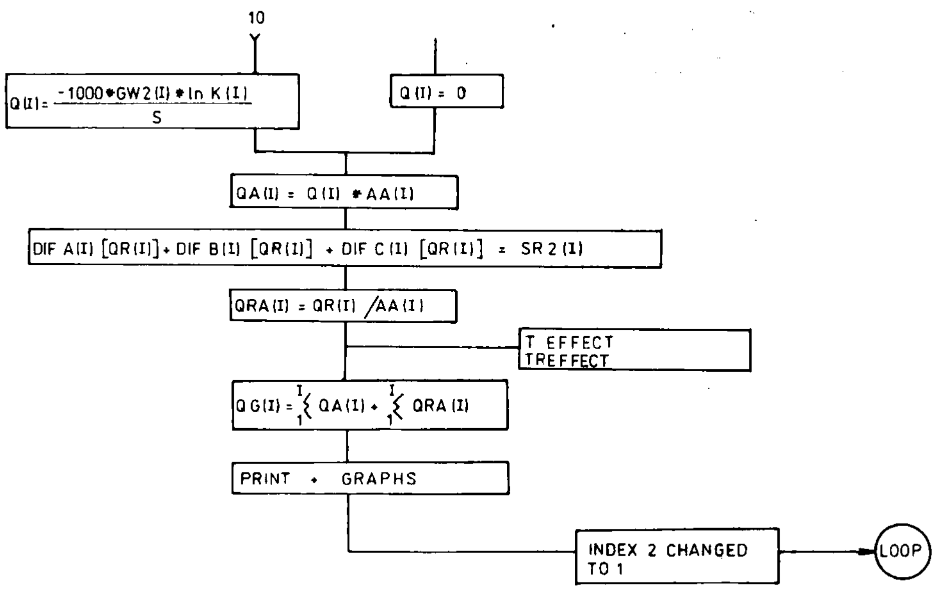
The capillary rise zone is replenished from the groundwater zone at the capillary rise rate $FFF(I)$, simulated in our version as a constant value. The zone of capillary rise is replenished until it becomes fully saturated. The zone can be emptied through evaporation.

Groundwater depletion is calculated as

$$Q(I) = \frac{-1000GW2(I)^* \ln K(I)}{S} \tag{6.14}$$

where $Q(I)$ is the yield, in $m^3 s^{-1} km^{-2}$; $GW2(I)$ is the groundwater storage at the end of the balanced interval, in mm; $K(I)$ is the recession coefficient; and S is the number of seconds in the balanced interval.

For $GW2(I)$ equal to or less than zero, an intermittent régime is simulated. A zero or negative value of $GW1(I)$ can occur as a result of prolonged groundwater outflow and transpiration from the zone of capillary rise without any replenishment. Such a situation is typical of arid and semi-arid regions.



6.10 Flow chart of routing. Partial flow chart 11.

Total baseflow is calculated by multiplying the yield by the drainage area of the subregion (Fig. 6.10).

The differential equation which is used to calculate the flood yield $QR(I)$ in a given time interval and for flood routing in the intervals that follow, is mentioned in subsection 6.4. Total flood discharge is again determined by multiplying the yield by the drainage area of the subregion.

Finally, the total discharge is calculated as the sum of baseflow and flood discharge from all the subregions.

If necessary the effect of the occurrence of $QA(I)$ and $QR(I)$ can be delayed by the parameters $T(I)$ and $TR(I)$. In a normal run both $T(I)$ and $TR(I)$ are equal to 1, and thus an immediate occurrence of the baseflow and flood flow in the stream is expected. However, if a delayed effect has been proved experimentally, the lag of $T(I)$ and $TR(I)$ hours can be introduced independently for any subregion distant from the lowest cross-section of the balanced area.

Any of the calculated variables given in subsection 5.3 can be obtained in a graphical or numerical form. As has already been mentioned, instead of instant daily values of the water balance, it is more convenient to compute the difference between the initial water balance values and the water balance values obtained for each particular day for the interception zone, the zone of aeration, the zone of capillary rise, the groundwater zone, the snowpack zone, and the zone of capillary pores within the groundwater zone. For the precipitation, evaporation and transpiration components and the total surface runoff and total groundwater outflow, the sums for the first day of the model application are calculated. The water balance equation based on the symbols given in Tab. 6.1 has the following form:

$$SUMP(I) = SDEE(I) + SDSR2(I) + SDGR(I) + DUM2(I) + DIMA2(I) + \\ + DIMB2(I) + DGW2(I) + DSNO2(I) + SWW(I) \qquad 6.15$$

where

$$SDEE(I) = SUMEP(I) + SUMETTP(I) + SUMETS(I) \qquad 6.16$$

6.12 MODEL TESTING

It is typical of a conceptual approach that many parameters enter the model and only some of them can be initially determined with a reasonable accuracy. Thus it cannot be expected that the first few runs will provide reliable results. In order to determine whether the model is working properly, some observed data should be available, which can also be calculated by the model. Usually, the most accessible are the groundwater level data and/or discharge data. A day-by-day comparison of one or both of these values with the calculated ones can be used to test the accuracy of the model. The test can be arranged so that for each time interval an accumulated standard error of estimate is calculated:

$$SH = \sqrt{\frac{\sum_{1}^{n}[HO(I) - HC(I)]^2}{n}} \qquad 6.17$$

and

$$SQ = \sqrt{\frac{\sum_{1}^{n}[QO(I) - QC(I)]^2}{n}} \qquad 6.18$$

where the "C" values are those calculated by the model and "O" values the observed ones; n is the number of events involved.

By comparing the final values from two runs with different parameters, we can judge whether the model has improved through the changing of one parameter. In the initial stage it is recommended that no more than one parameter be changed, otherwise the trial and error approach may not be successful.

Although several attempts were made to apply optimization methods, the trial and error method still remains the most convenient.

Even if the conceptual approach represents certain physical features, we should bear in mind that any concept of the water balance simulation is a compromise between a tank model on one hand and the reality on the other.

6.13 MODEL APPLICATION

To demonstrate the model application, the example of a forest catchment at Hartvíkov, Czechoslovakia is given. The catchment covers 0.984 sq.km, and about 10% of it is considered to be an area of fast response while the rest is part of the so-called deep region (0.886 sq.km). Because of the small size of the catchment, the time of travel of groundwater and surface water has been estimated at 1 hour only. According to the information in Fig. 6.3, $DIFB(I)$ was estimated at 2.6 and $DIFC(I)$ at 0.5 for both subcatchments. Porosities were measured in the field and found to be 0.160, 0.360, 0.4 and 0.165, 0.235 and 0.17 for noncapillary, capillary and wilting pores of deep and shallow subcatchments. AF and BF were estimated at 0.25 and 25.0 for both subcatchments respectively. Since a coniferous forest forms the dominant cover of the catchment, the difference between the maximum and minimum interception storage is small, 65 – 60 mm. Similarly, interception rates are less variable seasonally (5 – 4.4 mm hour^{-1} for the deep and 4.6 – 4.3 for the shallow subcatchment). The depth of roots throughout the years is 300 mm. Day 210 was found to be the maximum vegetational development day.

The infiltration rate $FFX(I)$ was estimated at 1 mm hour^{-1}, and the water release from interception $FMX(I)$ also at 1 mm hour^{-1} for both catchments. The rate of capillary rise was estimated at 0.04 mm hour^{-1}. The recession coefficient for the deep subcatchment was calculated as 0.9999, and for the shallow one 0.9989.

The deep subcatchment outflow existed only when the outflow from the shallow catchment ceased. A zero outflow from a catchment indicates that the rate of groundwater storage is also equal to zero, or it may be negative. An explanation for this is that after the groundwater level, and consequently the groundwater storage, has dropped to a certain critical level, groundwater outflow stops. However, the groundwater level and storage can be depleted through water uptake by transpiration. Thus is was assumed that a negative groundwater storage may appear in the calculation. An initial groundwater outflow from the deep subcatchment was 0.0015 m^3 s^{-1} and the yield was

$$Q(1) = \frac{0.0015}{0.8860} = 0.0017 \qquad \mathrm{m^3\,s^{-1}\,km^{-2}}$$

Then

$$Q(1) = \frac{-1000GW1(1)\ln K(1)}{S} \qquad \mathrm{m^3\,s^{-1}\,km^{-2}}$$

and

$$GW1(1) = \frac{Q(1)\,S}{-1000\ln K(1)} = \frac{0.0017\cdot 3600}{-1000\ln 0.9999} = 47\ \mathrm{mm}$$

For observed $HO(1)$

$$HO(1) = \frac{SG(1) - GW(1)}{NPOR(1)}$$

and thence

$$SG(1) = HO(1)\cdot NPOR(1) + GW(1) = 770\cdot 0.166 + 47 = 174.82\ \mathrm{mm}$$

Total soil moisture zone capacity is

$$SL(1) = HC(1)\,FPOR(1) = 770\cdot 0.36 = 277\ \mathrm{mm}.$$

Because

$SLB(1)$ was estimated by 100 mm,
$SLA2(1) = SL(1) - 100 = 177.2$ mm.
$ILMA1(1)$ was estimated at 88.6% of $SLA2 \sim 157$ mm

For $NAR = 2$ (shallow region) we expect that $Q(2) = 0$ which means that $IGW1(2) = 0$.

Thus

$SG(2) = HO(2)\,NPOR(2) = 2000 \cdot 0.165 = 330\,\mathrm{mm}.$
$SL(2) = 2000 \cdot 0.235 = 470\,\mathrm{mm}.$ Again for $SLB(2) = 100$:
$SLA2(2) = SL(2) - 100 = 470 - 100 = 370\,\mathrm{mm}.$
$ILMA1(2)$ was estimated at 96.4% of $SLA2(2) = 357\,\mathrm{mm}.$

6.14 REFERENCES

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7 STATISTICAL AND PROBABILITY ANALYSIS

7.1 GENERAL

There are many processes in groundwater hydrology for which the physical background is not fully understood, and cannot be understood except by applying the probability theory. For example, fluctuations in the groundwater level and in water discharged by springs and changes in quality of subsurface water, are all processes which have been observed over regular intervals for many years. One of the main purposes of modelling hydrological sequences of this type is to generate synthetic sequences which can be evaluated by probabilistic routines and used for forecasting.

Contrary to the conceptual models discussed in Chapter 6, these models are characterized as probabilistic or stochastic. As stated by Ven Te Chow [1] "... a stochastic process is different from a probabilistic process as the former is generally considered as time-dependent and the latter time-independent". Thus stochastic models are expected to reproduce the statistical characteristics of those historical series relevant to a given problem.

In contrast to probability, statistics deals with the processing of sampled data. The collection of all phenomena under consideration, called a population, is replaced by a segment and it is assumed that such a segment may have some characteristics associated with the population. An individual observation of any variable X, say groundwater level, is known as a variate. The observation of groundwater levels for a certain period of time is referred to as a trial, while the collection of all possible groundwater levels is called sample space.

The groundwater data obtained from a trial are a random variable which can be considered continuous, whereas the number of rainy days which may have contributed to the formation of the groundwater level is looked upon as a discrete random variable.

It is assumed in this Chapter that the meaning of basic statistical parameters (such as the mean, the median, the mode), together with the measures of variability (such as the mean deviation, the standard deviation, the variance, the range and the coefficient of variation), and also the measures of skewness, are generally known. Therefore the Chapter is directed towards some practical applications of probabilistic methods in the various fields of groundwater resources assessment.

There have been many attempts at time series analysis which have resulted in model solution and authors have usually concentrated on probabilistic and stochastic models. Of course, there are some disadvantages in applying stochastic models to processes whose physical background is not known. Sometimes the results can even be misused to support initial presumptions. Fortunately, the present and probable future development in this field is directed more towards searching for the linkage between stochastic and conceptual simulation.

7.2 TIME SERIES

A time series is a set of observations taken at specified times, usually at equal intervals. A time series is thus defined by the values $X(t_1)$, $X(t_2)$, $X(t_3)$ where $t_1 < t_2 < t_3$. In hydrological literature the term "hydrological sequence" is often used instead of time series.

Obviously many hydrological and geophysical processes can be considered as time series. The following factors can be considered as components of time series:

1. Long-term secular movements in a general direction in which the time series appears to extend over a long period of time. This type of secular movement is also called secular variation or secular trend, and in graphical representations is indicated by a trend curve. Groundwater level decline caused by long-term continuous pumping is a typical example of a secular movement.

2. Cyclical movements or cyclical variations refer to a long-term oscillation. They may or may not be periodic. For instance, in some time series, the effects of sunspots may develop cyclical variations in hydrological régimes (Anděl, Balek, Verner [2]).

3. Seasonal movements refer to variations of time series due to the effects of flood. Some of these impacts produce variations only for a short time; however, it is conceivable that they may become intense and result in other types of variations.

The random variable has a certain probability distribution. If this distribution remains constant, the process and the time series are considered stationary. For instance, groundwater fluctuation in a completely natural environment can be viewed as a stationary process. Man's activity and groundwater exploitation usually develop a non-stationary process. While in the former case the statistical parameters will remain unchanged for each segment of the time series, in the latter they may vary from segment to segment.

It is necessary sometimes to distinguish between random and non-random processes. In a random process, the variates composing the time series are considered to be independent of each other. Obviously, many hydrological processes may be more correctly considered as non-random.

7.3 MULTIPLE REGRESSION

It is frequently necessary to find a relationship between two or more hydrological phenomena. For two related variables linear or curvilinear regression has been commonly applied. With the advent of computer-oriented analysis, multiple regression, linear or curvilinear, has become widely used. From the many types of programme available in computer libraries, it is recommended to use those through which the optimum combination of the variables involved is chosen according to the level of their preselected significance:

$$y = a_0 + a_1x_1 + a_2x_2 + \ldots, a_mx_m \tag{7.1}$$

where $x_1, x_2, \ldots, x_m$ are variables and a_i the relevant coefficients. A more general form is given for multiple non-linear regression:

$$y = f(x_1, x_2, \ldots, x_m) \tag{7.2}$$

However, it is always preferable to transform the relation into the more convenient linear form.

This method is recommended provided there are many gaps in the observed data, and their quality does not allow the application of some of the stochastic models.

As an example, an analysis of the régime of three springs at Jáchymov, Czechoslovakia can be given. The springs, utilised for medical treatment, indicated during the 1971 – 1981 period a high variability of discharge, temperature and Rn^{222} content. Multiple regressional analysis was applied in order to find out whether the springs located in a deep mine had been influenced by the changing climatic pattern, and whether the snowmelt on the slopes adjacent to the mine might have influenced their régimes.

This type of problem requires adaptation of the precipitation variable, so that as well as the monthly precipitation total the monthly snowmelt total could be alternatively introduced. For each of the three springs the mean monthly discharge, radon content, air temperature, and monthly precipitation and monthly snowmelt totals were considered independent variables.

The following relationships were formulated for the alternative with monthly rainfall:

$$Q_1 = 0.471 + 1.235T_1 + 1.374R_1; \qquad R = 0.794,\ S = 1.648$$
$$Q_2 = 1533.291 + 26.504R_2 + 28.513T_2; \qquad R = 0.794,\ S = 46.973$$
$$Q_3 = 27.943 + 1.508R_3; \qquad R = 0.745,\ S = 3.081$$

Q_1, Q_2, Q_3 is the spring discharge, in litres $\min^{-1}$; T_1, T_2, T_3 is the water temperature, in °C, R_1, R_2, R_3 is mean radon content, in $nC_i\ 1^{-1}$; all values given as monthly means. R is the multiple correlation coefficient:

$$R = \sqrt{1 - \frac{S_1^2}{s_1^2}} \tag{7.3}$$

In other words, R is expressed by the ratio of the standard deviation of estimated values divided by the standard deviation of the actual values of the dependent variable. Finally, S is the unbiassed standard deviation of residuals:

$$S = \sqrt{\frac{\sum_{1}^{N}(y - y')}{(N - m)}} \qquad 7.4$$

where y is the observed value and y' is the estimated one; N is the sample size and m is the number of the parameters.

The critical value of the multiple regression coefficient for a given number of observations and variables at 5% level of significance was 0.399. No impact of the precipitation and air temperature was detected.

In the alternative with snowmelt involved, the following relation was obtained for Q_1:

$$Q_1 = 36.435 + 0.017P_m - 2.051R_1; \qquad R = 0.722, \ S = 2.202$$

Here P_m is the snowmelt total in a month in mm.

To ascertain whether there is any relation between the spring régimes, correlation coefficients of a simple linear regression between pairs of discharges were calculated:

$Q_1 - Q_2$, $R = 0.569$
$Q_1 - Q_3$, $R = 0.470$
$Q_2 - Q_3$, $R = 0.939$

From the results obtained it was concluded that the effects of air temperature and precipitation on the spring régimes were insignificant. A relatively low value of the coefficient at P_m indicated that the influence of the snowmelt was of little importance. There is a relation between the fluctuation of the discharge and radon content; with the exception of Spring 3, the temperature is also significant. There is a close relationship between Springs 2 and 3, while Spring 1 appears to have a régime of its own.

7.4 ANALYSIS OF PERIODICITY

As has been mentioned earlier, hydrological sequences are composed of a trend, one or more periodic components and a random variable, though one or more of these may not be present. In most cases, the trend can be described analytically or at least estimated, and if the parameters of the function are not known the method of least squares can be used to estimate them. An undamped periodic component, or a combination of these, may be represented by a summation of superimposed harmonics:

$$P_t = \sum_{i=1}^{s} (a_i \cos \omega_i t + b \sin \omega_i t) \quad 7.5$$

where $a_1, a_2, \ldots, a_s$ are real numbers, s is the natural number; and the frequencies ω_i are in the interval 0 to π. The frequency $\omega_k\,(1 \leq k \leq s)$ corresponds to a period of length

$$T_k = \frac{2\pi}{\omega_k} \quad 7.6$$

The random component is postulated as having zero mean. It may be represented as an identically distributed set of independent random variables. For the former case, an estimate of the variances can be provided, while for the latter a mathematical-statistical model can be constructed to characterize the behaviour of the random component.

If the values of the random component are $y_1 \ldots y_n$, where n is the number of members in the sequence, then let

$$y_t = a_1 y_{t-1} + a_2 y_{t-2} + \ldots a_r y_{t-r} + \varepsilon_t \quad 7.7$$

where $a_1 \ldots a_r$ are real numbers $(a_r \neq 0)$ and ε_t are independent random variables with zero means and equal variances σ^2. The representation given by equation 7.7 is called an autoregressive sequence of the r^{th} order. If all the real and complex roots of the sequence

$$\lambda^r - \alpha_1 \lambda^{r-1} - \alpha_2 \lambda^{r-2} \ldots \alpha_r = 0 \quad 7.8$$

have absolute value less than unity, then the autoregressive sequence is stationary. If the absolute value of one or more of the roots of Eq. 7.4 is greater than unity, then the sequence is called evolutive or explosive. However, according to Wise [5] the stationarity of the autoregressive scheme can be recognized directly from the coefficients $a_i \ldots a_r$ and it is thus not necessary to solve Eq. 7.8.

If a model as described by Eq. 7.7 is to be applied, then the order of the autoregressive scheme has to be derived first, then the autoregressive coefficients, and finally an estimate of the residual variance provided.

For the purpose of extrapolation, the trend and the periodic components, if present, may be extrapolated directly while the extrapolation $\hat{y}_{N+1}$ of the random component may be calculated using the formula

$$\hat{y}_{N+1} = a_1 y_N + a_2 y_{N-1} + \ldots a_r y_{N-r+1} \quad 7.9$$

which is valid for one step of extrapolation. For further extrapolation, the method described by Yaglom may be applied [3].

Let the discrete time series be denoted as $X_1 \dots X_n$ where N is an odd number such that

$$N = 2M + 1$$

and M here is a natural number. If this condition is not satisfied, the first number of the sequence may be eliminated. If it appears on a logical analysis that the trend component is a constant value then the members of the sequence fluctuate periodically, randomly or as a combination of both forms about a certain mean value, the estimate of which is the arithmetic mean:

$$\overline{X} = \frac{1}{N}\sum_{i=1}^{N} \overline{X}_i \qquad 7.10$$

By eliminating the constant trend value we obtain a sequence

$$x_t = \overline{X}_t - \overline{X} \qquad 1 \leq t \leq N$$

with mean zero.

It now remains to determine whether that sequence is composed of independently distributed random variables, or whether further analysis of the sequence must be undertaken. From the series of various tests described by Hannan the test based on the circular correlation coefficient [4] gives

$$R = \frac{\sum_{t=1}^{N} x_t x_{t+1} - N x^2}{\sum_{t=1}^{N} x_t^2 - N X^2} \qquad 7.11$$

(where $x_{N+1} = x_1$) was chosen. A treatise of the circular correlation coefficient together with a table of critical values was given by R. L. Anderson [7].

The test for the existence of a periodic component was provided by Fischer's test, suitably modified where necessary. Further extensions of the test have been described by Whittle [6] and Hannan [4].

The original Fisher's test is applicable when a given sequence is a combination of a periodic component and an independent random variable; the modified test is applied when the random variables are dependent. From our experience the modified test has not proved to be sufficiently sensitive; only its use in conjunction with an extremely long sequence contributes to increased sensitivity.

In accordance with Fisher's test the quantities

$$C_k = \frac{1}{N}\sum_{t=1}^{N-k} x_{t+k} x_t, \qquad k = 0, 1, \dots N-1 \qquad 7.12$$

are calculated.

Some authors call the function

$$B(k) = \frac{C(k)}{C(0)}, \qquad k = 0, 1, \dots N - 1$$

the autocorrelation function of the sequence $x_1 \dots x_n$.

The periodogram is defined as

$$I(\lambda) = \frac{1}{2\pi}\left[C_0 + 2\sum_{t=1}^{N-1} C_t \cos t\lambda\right] \qquad 0 \leq \lambda \leq \Pi \qquad 7.13$$

Put $\lambda_k = 2\Pi k/N$; $k = 1, 2, \dots M$, where M has been defined. Denote $I(\lambda_q) = \max [I(\lambda_1), I(\lambda_2), \dots I(\lambda_M)]$ and

$$g = \frac{I(\lambda_q)}{I(\lambda_k)}$$

If there is no periodic component of the type $a \cos \lambda t + b \sin \lambda t (= a' \cos(\lambda_t + \beta))$ present, and the sequence is composed of independent normal variables with equal means and variances, then the distribution of g is given by

$$G(x) = P(g > x) = \sum_{j=1}^{[1/x]}(-1)^{j-1}\binom{M}{j}(1 - jx)^{M-1} \qquad 7.14$$

where $[1/x]$ denotes the integral part of the number $1/x$.

Therefore, when $G(g) < \alpha$ the hypothesis on the absence of the periodic component can be rejected at the level α. In such a case another test may be provided on the presence of a further periodic component. According to Whittle, λ_q may be eliminated and new values found:

$$I(\lambda'_q) = \max[I(\lambda_1) \dots I(\lambda_{q-1}), I(\lambda_{q+1}) \dots I(\lambda_M)] \qquad 7.15$$

and

$$g' = \frac{I(\lambda'_q)}{\sum_{k=1}^{M} I(\lambda_k)}$$

The significance of g' is then tested by equation 7.15 where the value of M has been replaced by $(M - 1)$. This procedure is then repeated until significant results have been obtained.

Suppose that the procedure was applied s times ($s \geq i$) which means that s values from $\lambda_1 \dots \lambda_M$ were found to be significant. Then the parameters a_i, b_i of the periodic component P_i are calculated by the method of least squares, i.e. the values of

$$\sum_{t=1}^{N}[X_t - \sum_{i=1}^{s}(a_i \cos \omega_i t + b_i \sin \omega_i t)]^2$$

should be minimal.

If the deviations from the periodic function are dependent the following modification of Fisher's test is suggested. First, an estimate of the spectral density of the sequence $X_1 \ldots X_N$ is provided by Parzen's formula:

$$f(\delta_j) = \frac{C_0}{2\pi} + \frac{1}{\pi}\sum_{k=1}^{n/2}\left[1 - \left(1 - \frac{k}{n}\right)\frac{6k^2}{n^2}\right] C_k \cos\frac{\Pi k_j}{n} +$$

$$+ \frac{2}{\Pi}\sum_{k=n/2}^{n}\left(1 - \frac{k}{n}\right)^3 C_k \cos\frac{\Pi k_j}{n} \qquad 7.16$$

where

$$\delta_j = \frac{\Pi_j}{n} \qquad j = 0, 1, \ldots n,$$

and n is an even number chosen between $N/6$ and $N/5$. Linear interpolation of the values $f(\delta_j)$ allows f to be evaluated over the whole interval. To test for the existence of a periodicity, further quantities are evaluated

$$K_j = \frac{I(\lambda_j)}{f(\lambda_j)} \qquad 1 \leq j \leq M$$

$$k_q = \max(k_1, \ldots k_M)$$

and

$$g = \frac{k_q}{\sum_1^M k_j}$$

is calculated.

The value of g is then tested for significance by using equation 7.14 in the original test.

If the presence of a periodic component has been verified, the random variables $Y_t = X_t - P_t$ are analysed. Otherwise $P_t = 0$ and $y_t = x_t$ and the test for independence is provided by Anderson's test, based on the circular correlation coefficient calculated from equation 7.11, where x_t is replaced by y_t. According to Anderson [7] the determination of the significance of the circular correlation coefficient should be slightly modified when the deviation from the periodic component are analysed. For the case where $P_t = 0$ the circular correlation coefficient of the $y_t(=x_t)$ already has been calculated.

The order of the autoregressive model 7.7 is determined by Whittle's test [6]. For its application denote

$$D_k = \frac{1}{N}\sum_{t=1}^{N-k}(y_{t+k} - \bar{y})(y_t - \bar{y}), \qquad k = 0, 1, \ldots N - 1 \qquad 7.17$$

where

$$\bar{y} = \frac{1}{N}\sum_{1}^{N} y_i$$

If a periodic component has not been detected then

$$D_k = C_k, \qquad 0 \le k \le N - 1$$

where C_k are the values calculated from equation 7.12.

Then the determinants

$$\Delta_p = \begin{vmatrix} C_0, & C_1 & \dots & C_p \\ C_1, & C_0 & \dots & C_{p-1} \\ C_p, & C_{p-1} & \dots & C_0 \end{vmatrix} \qquad p = 0, 1, \dots 30 \qquad 7.18$$

are calculated.

Bearing in mind the minimum sequence length available, an upper limit of 30 is placed on the order of the autoregressive schemes considered, in order to preserve the statistical validity of the results. Then the values

$$\lambda_p = \frac{\Delta_{p+1}\Delta_{p-1}}{\Delta_p^2} \qquad 1 \le p \le 29$$

$$\Psi_p^2 = (N - P - 1)\frac{1 - \lambda_p}{\lambda_p} \qquad 1 \le p \le 29$$

are calculated.

If the tested sequence is autoregressive of order p then Ψ_p^2 is asymptotically distributed as χ^2 with 1 degree of freedom. Thus Ψ_p^2 can be used to test the hypothesis that the sequence $y_1 \dots y_N$ is autoregressive of order p against the alternative hypothesis that it is of order $p + 1$.

A further test due to Whittle used in the analysis is based on the following principle: when

$$\hat{U}_k = \frac{N\Delta_k}{k - 1}, \qquad k = 1, 2, \dots 10,$$

then

$$\Psi_p^2 = (N - p - q)\frac{\hat{U}_p - \hat{U}_{p+q}}{\hat{U}_{p+q}}$$

is asymptotically distributed as χ^2 with q degrees of freedom. The value of Ψ_p^2 then tests the hypothesis that the autoregression of the order p is significant as opposed to the autoregression of order $p + q$. The values Ψ_p^2 (starting with $p = 1$) are compared with the corresponding critical values of the Chi-square distribution with q degrees of freedom. The coefficients $a_i \dots a_r$ in the model are estimated by the method of least squares such that

$$\sum_{t=r+1}^{N} (y - a_1 y_{t-1} \dots a_r y_{t-r})^2 \qquad 7.19$$

is minimised. Mann and Wald [8] proved that such estimates are consistent and have asymptotic normal distribution. Even if the value of Ψ_1^2 is not significant, an autoregressive model of the first order is evaluated. The application of it depends on a complex analysis of the results, and particularly on the significance of the circular correlation R of the sequence $Y_1 \dots Y_N$.

The quantities

$$\hat{\sigma}_p^2 = \frac{\Delta_p}{\Delta_{p-1}} \qquad 1 \leq p \leq 29$$

are now calculated. If the autoregression of order r has been found significant then σ_r^2 is an estimate of the variance σ^2 of the component ε_t in the model 7.7. Finally, the quantity

$$S^2 = \frac{1}{N} \sum_{t=1}^{N} y_t^2$$

is calculated and used as an estimate of the variance of the random variable y_t. On completion of the foregoing analysis, a further analysis by what is known as the method of hidden periodicities can be provided. The theoretical form of the periodogram may be defined as

$$I_N(\lambda) = \left| \frac{1}{2\Pi N} \sum_{t=1}^{N} x_t \, e^{-it\lambda} \right|^2 \qquad 7.20$$

If the sequence $x_i \dots x_N$ contains a component

$$a \cos(\lambda_0' t + \beta) + y_t, \qquad 0 < \lambda_0 < \Pi, \; a \neq 0 \qquad 7.21$$

such that

$$x_t = a \cos(\lambda_0' t + \beta) + y_t \qquad 7.22$$

where $y_t \dots y_N$ are either independent or dependent random variables, then the approximation

$$I_N(\lambda_0') = \frac{a^2 N}{8\Pi} \qquad 7.23$$

is valid.

If $y_1 \dots y_N$ can be considered as independent random variables with zero means and variances σ^2, then equation 7.23 is more accurate for larger values of N, this being equally true for a given N but smaller σ. Equation 7.23 gives the value of the periodogram at the peak corresponding to the frequency λ_0'. As already cited, the basic criterion for periodogram analysis is Fisher's test based on the value

$$g = \frac{I_{\max}}{\sum_1^m I_k} \qquad 7.24$$

where

$$I_k = I_N(\lambda_k), \qquad \lambda_k = \frac{2\Pi k}{N}, \qquad 1 \le k \le M, \qquad M = N - \frac{1}{2}$$

The distribution of g is given by equation 7.14.

A more convenient form of calculation is given as

$$P(g > x) \doteq M(1 - x)^{M-1} \qquad 7.25$$

Fisher's test is applicable in the case where only one periodicity is contained in a sequence such as that given by Eq. 7.22. With an increase in the number of periodicities, the power of the test decreases rapidly. In order to detect several periodic components of the type postulated in Eq. 7.22 a much longer realization is necessary. Suppose that the sequence contains only one periodic component of the type postulated in Eq. 7.22 and that the peak of the periodogram denoted by K corresponds with it. If

$$\sum_{k=1}^{m} I_k - K = G \qquad 7.26$$

then

$$g = \frac{K}{K + G}$$

Suppose that for instance this periodicity has been detected by Fisher's test at the 2% level which means that

$$P\left\{g \ge \frac{K}{K + G}\right\} = 0.02 \qquad 7.27$$

An example was studied with $N = 75$ and $M = 37$.
According to equation 7.25 it can be shown that Eq. 7.27 is valid for

$$\frac{K}{K + G} \doteq 0.189 = b \qquad 7.28$$

Now another situation can be tested where the sequence contains two periodicities of the type $a \cos(\lambda'_0 t + \beta_0)$ and $a \cos(\lambda'_1 t + \beta_1)$ where $\lambda'_1 \neq \lambda'_0$. Then the periodogram produces two peaks, each of them approximately equal to K. The existence of a periodicity may be tested by g:

$$g = \frac{K}{2K + G}$$

because the value of G has been changed only slightly by omitting one of the values I_k. From Eq. 7.28.

$$K = \frac{bG}{1 - b}$$

is calculated and thus

$$\frac{K}{K + G} = \frac{b}{1 + b} \doteq 0.158\,96$$

According to Eq. 7.25.

$$P\left\{g > \frac{K}{2K + G}\right\} = P\{g > 0.158\,96\} \doteq 0.0739$$

Similarly, if these periodicities exist in the sequence of the form

$$a \cos(\lambda'_i t + \beta_i) \qquad 1 \leq i \leq 3\,, \qquad 0 \leq \lambda'_1 \neq \lambda'_2 \neq \lambda'_3 < \pi$$

then

$$g = \frac{K}{3K + G} = \frac{b}{1 + 2b} \doteq 0.137\,16$$

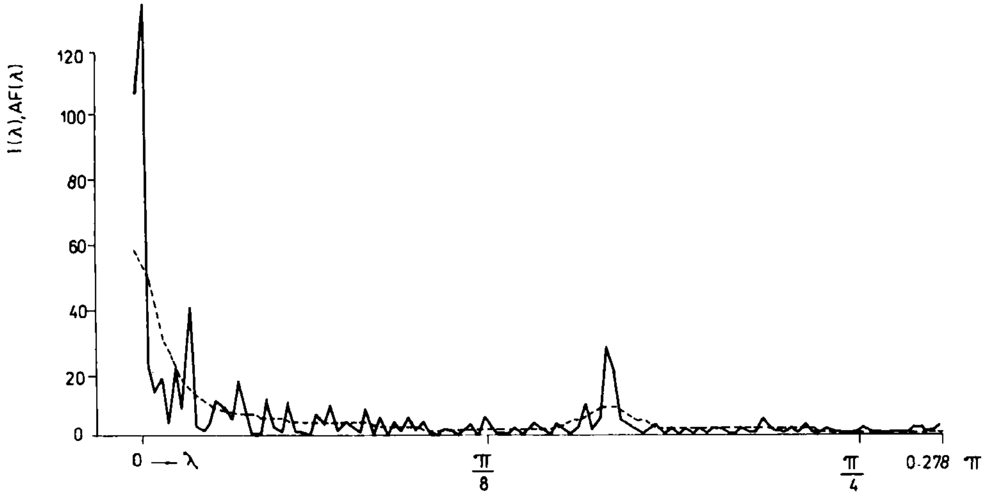
and

$$P\{g > 0.137\,16\} \doteq 0.1825$$

Thus in a case where two or three periodicities are in the sequence, Fisher's test does not verify the existence of even one of them at the 5% level of significance. For the case where a single periodicity exists, significance at the 5% level only would be obtained. Obviously the power of the test depends on the ratio of the amplitudes of the periodic components (Anděl, Balek [9]).

Programmes for the periodicity analysis can be found in computer libraries in more or less similar form, however, the test of significance of the periodicities is not usually performed in such programmes, and the periodogram analysis has to be done visually by selecting the most pronounced peaks of the periodogram.

In Fig. 7.1 is an example of the periodogram of mean monthly discharges of the Sadový spring, one of the well-known springs at Karlovy Vary, Czechoslovakia. The spring has been observed since August 1908, and the relatively few missing data were calculated by correlating the discharges with the discharges of other springs located in its vicinity. In the same Figure also the spectral density graph is plotted. As can be traced in the periodogram, the periodical components for frequencies 0.015, 0.0679, 0.128, 0.052 78 correspond (on a monthly basis) to 34.7, 7.71, 4.08 and 9.92 years respectively.



7.1 Periodogram of mean monthly discharges of the Sadový spring. The most significant frequencies are 0.015, 0.0679, 0.128, 0.052 78 and 0.524.

A joint evaluation of the periodical components of the discharge sequences is in Tab. 7.1. The same Table also shows periodical components in the temperature, rainfall, air pressure, and evaporation residue sequences. Through such analysis the relationship between spring régimes can be determined. For example, we can see that the period of 34 years, probably also related to the period of 17 years, was found in the majority of the sequences. Perhaps the period of 8 – 9 years also can be considered as related to them. Thus from a given Table we can conclude that the Vřídlo, Tržní, Sadový, Libušin, Mlýnský and Svobody springs have similar régimes.

As far as the water temperature is concerned, similar régimes can be traced in springs Tržní, Zámecký, Skalní, Sadový, Libušin, Mlýnský and Svobody springs.

Another significant period in the analysed sequences is 5 – 6 years (probably related to the period 11 – 12 years) found for the Vřídlo, Tereza, Václav, Zámecký, Libušin and Mlýnský springs.

A significant frequency of 0.524 corresponds to a one-year period, which indicates that the seasonal component is significant. Quite possibly all the springs are partly influenced by the climate.

Tab. 7.1 The most significant periodicities in the analyzed sequences

Spring	Observed	Periods in order of importance			
Vřídlo, discharge	1870 – 1978	36.01	10.80	15.43	4.70
Tereza, discharge	1908 – 1977	23.11	13.78	8.67	5.33
Tržní, discharge	1908 – 1977	34.70	17.35	9.92	6.94
Zámecký, discharge	1908 – 1977	17.35	34.07	6.31	7.71
Skalní, discharge	1908 – 1977	17.35	6.31	–	–
Sadový, discharge	1908 – 1977	34.70	7.71	4.08	9.92
Václav, discharge	1921 – 1977	18.97	7.11	5.17	3.56
Libušin, discharge	1908 – 1977	34.70	11.57	6.31	4.34
Mlýnský, discharge	1908 – 1977	34.68	11.58	8.67	5.78
Svobody, discharge	1908 – 1977	17.35	8.68	7.71	–
Tereza, temperature	1908 – 1977	17.40	9.92	1.01	6.31
Tržní, temperature	1908 – 1977	17.35	1.01	4.34	8.68
Zámecký, temperature	1908 – 1977	34.70	17.35	7.71	4.96
Skalní, temperature	1908 – 1977	17.35	1.05	11.57	7.71
Sadový, temperature	1908 – 1977	34.70	7.71	9.92	1.01
Václav, temperature	1921 – 1977	18.97	9.39	1.00	5.17
Libušin, temperature	1908 – 1977	1.01	23.00	9.90	2.89
Mlýnský, temperature	1908 – 1977	1.00	17.33	11.56	6.93
Svobody, temperature	1908 – 1977	17.35	34.70	11.57	1.01
Vřídlo, evap. residue	1910 – 1934	6.08	3.48	1.62	0.52
Vřídlo, chloride	1910 – 1934	6.08	3.48	2.43	0.67
Vřídlo, sulphur	1910 – 1934	–	–	–	–
Climate					
Karlovy Vary, air temp.	1908 – 1977	7.11	1.00	–	–
Karlovy Vary, precipit.	1908 – 1977	1.00	0.50	–	–
Karlovy Vary, air press.	1908 – 1977	0.99	4.96	0.33	13.88

7.5 COHERENCE OF TIME SERIES

In hydrological studies it is often necessary to analyse the relationship of two hydrological phenomena. Such an analysis may help to trace mutual interrelations, or at least to give a measure of similarity between two or more investigated sequences. The correlation coefficient is frequently used for this purpose. A natural generalisation of the correlation coefficient is the cross-correlation function (Anděl, Balek [10]). Unfortunately the statistical properties of this function are very complicated. Therefore a method based on the concept of coherence coefficients is applied. Let us consider two real time series

$$\ldots, x_{-1}, x_0, x_1, \ldots$$
$$\ldots, y_{-1}, y_0, y_1, \ldots$$

and we shall assume that the series are normally distributed. Many hydrological sequence series have constant but non-zero mean value. Substracting the empirical mean value we obtain a new series with zero expectation and approximately the same covariance structure. The function

$$R_{xx}(s,t) = Ex_s x_t \tag{7.30}$$

is called the covariance function of the series $\{x_t\}$. Similarly, $R_{yy}(s,t) = Ey_s y_t$ is the covariance function of the series $\{y_t\}$. These functions are sometimes called autocovariance functions. Two other functions

$$R_{xy}(s,t) = Ex_s y_t \quad \text{and} \quad R_{yx}(s,t) = Ey_s x_t \tag{7.31}$$

are the cross-covariance functions. If all four above mentioned functions depend only on the difference of their variables, i.e. $R_{xx}(s,t) = R_{xx}(s-t)$ etc., the two-dimensional time series is called stationary. Under certain conditions it is possible to express these covariance functions in a stationary case, using spectral density function in a spectral form. If

$$\sum_{t=-\infty}^{\infty} |R_{xx}(t)| < \infty\,, \tag{7.32}$$

then there exists a function $f_{xx}(\lambda)$, $-\pi \leq \lambda \leq \pi$ such that

$$R_{xx}(t) = \int_{-\pi}^{\pi} e^{it} f_{xx}(\lambda)\, d\lambda \qquad \text{for} \qquad t = \ldots,\ -1, 0, 1, \ldots \tag{7.33}$$

If conditions similar to the above are satisfied for R_{xy}, R_{yx} and R_{yy}, we can write them in the same form using other functions $f_{xy}(\lambda)$, $f_{yx}(\lambda)$ and $f_{yy}(\lambda)$. The functions f_{xx} and f_{yy} are the spectral density functions and f_{xy} and f_{yx} are called cross-spectral density functions. The functions $f_{xx}(\lambda)$ and $f_{yy}(\lambda)$ are generally complex functions satisfying $f_{xy}(\lambda) = f_{xy}(-\lambda)$, $f^*_{yx}(\lambda) = f_{yx}(-\lambda)$ and $f_{xy}(\lambda) = f^*_{yx}(\lambda)$, where the asterix denotes complex conjugation.

It is generally known that the stationary series x_t and y_t can be written in the form

$$x_t = \int_{-\pi}^{\pi} e^{it\lambda}\, dZ_x(\lambda)\,, \qquad y_t = \int_{-\pi}^{\pi} e^{it\lambda}\, dZ_y(\lambda)\,, \tag{7.34}$$

where $Z_x(.)$ and $Z_y(.)$ are random measures corresponding in some sense to the spectral density functions $f_{xx}(\lambda)$ and $f_{yy}(\lambda)$. Roughly speaking, x_t and y_t can be regarded as the sums of many functions $e^{it\lambda}$ with random coefficients. After some rearranging we can look upon them as if they were sums of functions $A_\lambda \cos(\lambda t + B_\lambda)$, where A_λ and B_λ are random variables. The natural question is whether the cosine components of the two series are independent or not. The degree of dependence does not need to be the same for different frequencies λ. As a measure of dependence of these components the coherence coefficient $C_{xy}(\lambda)$ was suggested. It is defined by the formula

$$C_{xy}(\lambda) = |f_{xy}(\lambda)|\,[f_{xx}(\lambda)f_{yy}(\lambda)]^{-1/2}, \qquad \text{if} \qquad f_{xx}(\lambda)f_{yy}(\lambda) \neq 0, \tag{7.35}$$

otherwise $C_{xy}(\lambda) = 0$. Some authors call $C^2_{xy}(\lambda)$ the coherence coefficient rather than $C_{xy}(\lambda)$.

A basic property of the coherence coefficient is that

$$0 \leq C_{xy}(\lambda) \leq 1 \tag{7.36}$$

If the series $\{x_t\}$ and (y_t) are independent then $C_{xy}(\lambda) = 0$. The upper limit 1 is reached, for example, if y_t is the same as x_t or if y_t arises from x_t by a shift in time.

Gelfland and Yaglom proved that [11]

$$I = -(2\pi)^{-1}\int_{-\pi}^{\pi} \ln(1 - C_{xy}(\lambda)\,\mathrm{d}\lambda \tag{7.37}$$

is the amount of information (per unit of time) contained in the series (y_t) about (x_t). It gives a different view of the role of $C_{xy}(\lambda)$.

The function $f_{xy}(\lambda)$ is generally complex. Let us write it in the form

$$f_{xy}(\lambda) = c_{xy}(\lambda) + i\,q_{xy}(\lambda), \tag{7.38}$$

where $c_{xy}(\lambda)$ and $q_{xy}(\lambda)$ are real and imaginary parts of $f_{xy}(\lambda)$. Using them, we obtain another formula for evaluating the coherence coefficient:

$$C_{xy}(\lambda) = \sqrt{\left[\frac{c^2_{xy}(\lambda) + q^2_{xy}(\lambda)}{f_{xx}(\lambda)f_{yy}(\lambda)}\right]} \tag{7.39}$$

Define the phase function:

$$\Phi_{xy}(\lambda) = \operatorname{arctg}\frac{q_{xy}(\lambda)}{c_{xy}(\lambda)}, \tag{7.40}$$

which has the following meaning: if b is an integer and $y_t = x_{t-b}$, then

$$\Phi_{xy}(\lambda) = \operatorname{arctg}(\operatorname{tg} b\lambda). \tag{7.41}$$

In this case the graph of $\Phi_{xy}(\lambda)$ consists of several parallel segments, each of them having the slope b. Thus we are able to detect shift b from $\Phi_{xy}(\lambda)$.

Let us consider a more general model

$$y_t = \sum_{k=1}^{p} a_k x_{t-b_k} \tag{7.42}$$

where $a_1, \dots a_p$ are real numbers and $b_1, \dots b_p$ are integers. From the mathematical point of view the series $\{y_t\}$ arises from $\{x_t\}$ by filtering. It can be shown that

$$f_{xy}(\lambda) = \sum_{k=1}^{p} a_k \exp\{ib_k\lambda\} f_{xx}(\lambda), \qquad 7.43$$

so that

$$\frac{c_{xy}(\lambda)}{f_{xx}(\lambda)} = \sum_{k=1}^{p} a_k \cos b_k\lambda \qquad 7.44$$

and

$$\frac{q_{xy}(\lambda)}{f_{xx}(\lambda)} = \sum_{k=1}^{p} a_k \sin b_k\lambda \qquad 7.45$$

If y_t is generated by the above general model, then analysis of the above functions can yield some estimates for a_k and b_k $(k = 1, \dots p)$.

Usually we know only a part of the two series $x_1, \dots x_N$, $y_1, \dots y_N$. Let us denote

$$H_{xx}(t) = N^{-1} \sum_{s=1}^{N-t} x_{s+t} x_s, \qquad H_{yy}(t) = N^{-1} \sum_{s=1}^{N-t} y_{s+t} y_s, \qquad 7.46$$

$$H_{xy}(t) = N^{-1} \sum_{s=1}^{N-t} x_{s+t} y_s, \qquad H_{yx}(t) = N^{-1} \sum_{s=1}^{N-t} y_{s+t} x_s;$$

using Parzen's coefficients

$$\omega_t = \begin{cases} (2\pi)^{-1}[1 - 6t^2(1 - t/m)/m^2] & \text{for } t = 0, 1, \dots, m/2, \\ \pi^{-1}(1 - t/m)^3 & \text{for } t = m/2 + 1, \dots, m \end{cases} \qquad 7.47$$

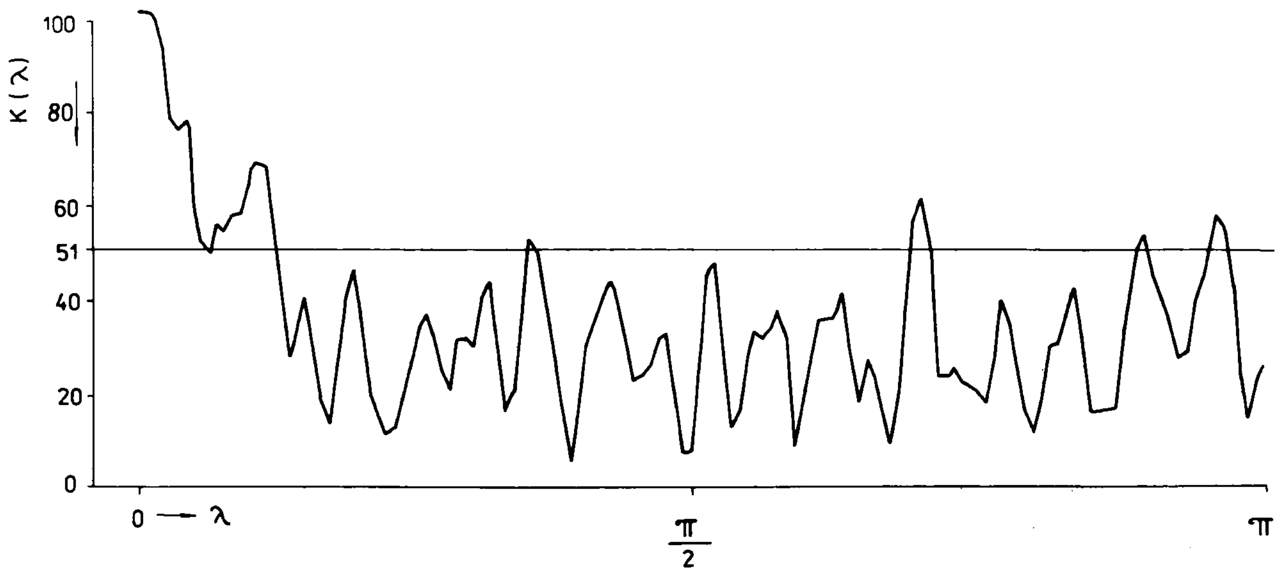
where m is an even integer usually chosen as $N/6$, approximately, we obtain estimates

$$f_{xx}(\lambda) = \omega_0 H_{xx}(0) + 2 \sum_{t=1}^{m} \omega_t H_{xx}(t) \cos t\lambda \qquad 7.48$$

$$f_{yy}(\lambda) = \omega_0 H_{yy}(0) + 2 \sum_{t=1}^{m} \omega_t H_{yy}(t) \cos t\lambda, \qquad 7.49$$

$$c_{xy}(\lambda) = \omega_0 H_{xy}(0) + \sum_{t=1}^{m} \omega_t [H_{xy}(t) + H_{yx}(t)] \cos t\lambda, \qquad 7.50$$

$$q_{xy}(\lambda) = \sum_{t=1}^{m} \omega_t [H_{yx}(t) - H_{xy}(t)] \sin t\lambda. \qquad 7.51$$



7.2 Coherence graph for the discharge of the Vřídlo and Sadový springs.

For the sake of simplicity, we use the same symbols for the estimates of the spectral density functions as for these functions themselves. The sample coherence coefficient $C_{xy}(\lambda)$ is then defined again by formula 7.39 and the sample phase function $\Phi_{xy}(\lambda)$ is defined analogously by 7.38. All the sample functions f_{xx}, f_{yy}, c_{xy}, q_{xy}, Φ_{xy} are computed at the points $\lambda_j = \pi j/m$ for $j = 0, 1, 2, \ldots m$.

The distribution of the sample coherence coefficient C_{xy} is given in tables published by Amos and Koopmans [12]. A short table is also reproduced in the book by Granger and Hatanaka [13]. If $C_{xy}(\lambda)$ exceeds the critical value, we reject the hypothesis that the theoretical coherence coefficient equals zero. In the above publication there is also a Table which can be used to construct the confidence interval of the theoretical phase function.

Using estimates of the spectral density function we can estimate the function given in formulae 7.44 and 7.45. We know their estimated values at the points $\lambda_j = \pi j/m$, for $j = 0, \ldots, m$. Their periodograms inform us about a_k and b_k. We compute the values of the periodograms at the points $\omega_k = \pi k/m$, $k = 0, 1, \ldots, m$. This is not the usual choice of points. (For Fisher's test we should compute, for example, the periodogram at points $2\pi k/(m + 1)$). Suppose that there is $a_k \neq 0$ for some k. For $\lambda = \lambda_j$ the corresponding component in 7.44 is

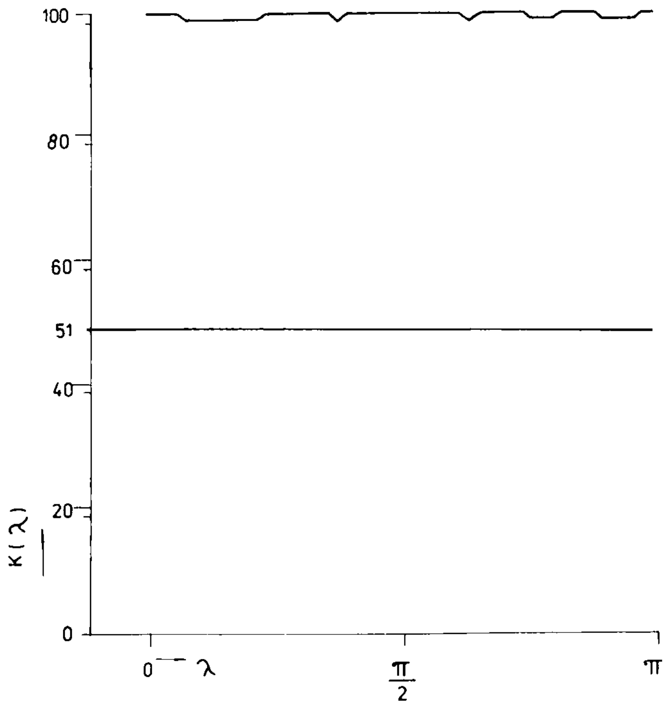
$$a_k \cos b_k \lambda_j = a_k \cos \left(\frac{b_k \pi}{m}\right) j .$$

The periodogram of $c_{xy}(\lambda)/f_{xy}(\lambda)$ must reach a high value at the point $b_k\pi/m$, i.e. at the point ω_{b_k}. The same holds true for the periodogram of $q_{xy}(\lambda)/f_{xy}(\lambda)$. It can be concluded that if the periodograms of $c_{xy}(\lambda)/f_{xy}(\lambda)$ and $q_{xy}(\lambda)/f_{xy}(\lambda)$ have high values at the point ω_h, that it indicates a component of the $a_k x_{t-b_k}$ type in 7.42, where $|b_k| = h$; b_k can be determined from hydrological analysis or alternatively from the phase function. Usually, if $\{x_t\}$ has some influence on $\{y_t\}$ and the influence of $\{y_t\}$ on $\{x_t\}$ is negligible, we have $b_k \geq 0$.

It happens very often in hydrological time series that $\{x_t\}$ and $\{y_t\}$ are highly correlated for the same t, however the correlation between x_{t+s} and y_t is not too high for $s \neq 0$. Then the periodogram of $c_{xy}(\lambda)/f_{xx}(\lambda)$ reaches a peak at the point 0 and has some influence on the points in the neighbourhood. For this reason, conclusions about $b_k \neq 0$ based on the periodogram of $q_{xy}(\lambda)/f_{xx}(\lambda)$ should be preferred in such a case.

The coherence graph for the discharge of Vřídlo and Sadový springs at Karlovy Vary can be given as an example (Fig. 7.2). Another example (Fig. 7.3) is provided for the discharge of Vřídlo and its evaporation residue. According to Amos and Koopmans, for a 5% significance level the critical value is 0.51. Thus in the former case a close relationship between both sequences corresponds to the frequencies 0.022 44 – 0.3815 and to the respective periods of 23 – 1 years. Other significant values were obtained for the seasonal periodical component of 6, 3 and 2 months.

In the latter case a significant coherence was obtained for all frequencies – an indication of a close relationship between both sequences.



7.3 Coherence graph for the discharge of Vřidlo and its evaporation residue.

7.6 DISTURBANCES IN HYDROLOGICAL SEQUENCES

Periodical components traced in time series are far from being stable. For many purposes the periodical components are expected to be stable, however, under various natural and artificial impacts the stability of the phase and amplitude can be easily disturbed. Thus the stability of the periodic components should always be examined, particularly if the theoretical approach has to be extrapolated to reach requested practical conclusions. In principle, usually the following questions have to be answered:

1. Are there any disturbances in the periodic components?
2. When are they most frequent?
3. How can their origin be explained?

If we are able to answer these questions, a physical interpretation of the statistical results becomes more realistic. For this purpose a model of disturbances in hydrological sequences was developed by Anděl and Balek [14].

For stationary sequences let ... X_{-1}, X_0, X_1,... be a random stationary sequence with zero mean. Put

$$Y_t = \sum_{k=-m}^{m} C_k X_{t-k}, \qquad t = \ldots, -1, 0, 1, \ldots \tag{7.52}$$

where $C_{-m}, \ldots C_m$ are given numbers. These numbers are called a filter. If $C_{-m} \neq 0$, $C_m \neq 0$, the filter is of a length $2m + 1$. The filters satisfying $C_k = C_{-k}$ for $k = 1, 2, \ldots, m$ are called symmetric or cosine filters. The filters of this type do not produce a phase change and therefore they are preferred. The function

$$h(x) = \sum_{k=-m}^{m} C_k \mathrm{e}^{-ikx} \tag{7.53}$$

is called the frequency response function; $|h(x)|^2$ is then called the power transfer function. If $\{X_t\}$ has a spectral density $f(\lambda)$, then the sequence $\{Y_t\}$ defined in 7.52 has the spectral density

$$f_y(\lambda) = |h(\lambda)|^2 f(\lambda) \tag{7.54}$$

A further filtering of the sequence $\{Y_t\}$, using a filter $d_{-n}, \ldots, d_n$ leads to a sequence $\{Z_t\}$ so that

$$Z_t = \sum_{k=-n}^{n} d_k Y_{t-k} \tag{7.55}$$

The spectral density of the sequence Z_t is

$$f_z(\lambda) = |k(\lambda)|^2 \, |h(\lambda)|^2 f(\lambda) \tag{7.56}$$

where

$$k(\lambda) = \sum_{k=-n}^{n} d_k \mathrm{e}^{-ik\lambda}$$

Very often a filter is required which transmits only frequencies from a certain band without change, while the others are suppressed. If we want $\{Y_t\}$ to have the spectral density $f(\lambda)$ for $\lambda \varepsilon(-\delta,\delta)$ and zero outside (where δ is a small positive number), then the power transfer function has to be 1 on $(-\delta,\delta)$ and 0 outside. Such a filter is called a low pass filter, because it transfers only low frequencies. Since any real hydrological sequences are limited to filters of rather short length, the mathematical requirements given above cannot be fully satisfied. We try to choose the coefficients C_k so that the filter may be a reasonable approximation of the originally required ideal filter. Sometimes the filtering is realized sequentially. Two or more filters are used and the final spectral density is given by formula 7.56. Since we are often restricted to filters of a very special type, a product $|k(\lambda)|^2\,|h(\lambda)|^2$ can be found a better approximation of the required function.

A moving average is known as a reasonably good low pass filter. Such a filter of length $2m + 1$ has the coefficients

$$C_k = \frac{1}{2m+1} \qquad \text{for} \quad k = -m, \ldots, m \tag{7.57}$$

Sometimes it is necessary to use the coefficients which are produced from two successive moving averages. This has to be done particularly when one moving average has been calculated from an even number of values. Then the other one must also have an even number of values, otherwise the final composite filter would not be symmetric. The advantage of filters produced by moving average is that the corresponding power transfer function vanishes at points which can be easily calculated.

A different approach is used for the nonstationary sequences. Let $\{X_t^0\}$ be a stationary random sequence with zero mean. It is well known that X_t^0 can be expressed in the form

$$X_t^0 = \int_{-\pi}^{\pi} \mathrm{e}^{it\lambda}\,\mathrm{d}Z(\lambda) \tag{7.58}$$

where $Z(\cdot)$ is a random measure. This measure corresponds to the spectral distribution function $F(\lambda)$ of the sequence in such a sense that

$$EZ(B_1)\,\overline{Z(B_2)} = \int_{B_1 \cap B_2} \mathrm{d}F(\lambda) \tag{7.59}$$

where B_1 and B_2 are Borel subsets of $(-\pi,\pi)$ and the bar denotes the complex conjugate. The physical meaning of formula 7.58 is that the stationary sequence $\{X_t^0\}$ is "a sum" of a great number of cosine curves with random amplitudes and random phase. The amplitudes and phases change randomly in different realizations of the sequence, but for a given realization they are constant.

The basic model of nonstationary sequences with zero means can be formulated as follows:

a) Let $a(t,\lambda)$ be a function, where $t = \ldots, -1, 0, 1, \ldots, \lambda \varepsilon < \pi,\pi >$.

b) The function $a(t,\cdot)$ is measurable for any t.

c) $a(t,\lambda) = \overline{a(t,-\lambda)}$.

d) $\int_{-\pi}^{\pi} |a(t,\lambda)|^2\,\mathrm{d}F(\lambda) < \infty$.

e) Put $X_t = \int_{-\pi}^{\pi} \mathrm{e}^{it\lambda}\,a(t,\lambda)\,\mathrm{d}Z(\lambda)$.

Then $\{X_t\}$ is a nonstationary sequence. It can again be considered as "a sum of cosine curves with random amplitudes and random phases", but now amplitudes and phases change in time even in any given realization.

If the function $a(.,\lambda)$ changes slowly in t, then the components of the sequence $\{X_t\}$ can be separated by filtering. The application of the filter is similar to the stationary case.

For the demodulation of nonstationary sequences, the investigation of the amplitude and the phase in time for a given frequency consists of the following steps:

a) Two new sequences $\{P_t\}$ and $\{Q_t\}$ are calculated from given (nonstationary) series X_t with zero mean by the formulae

$$P_t = X_t \sin \omega t$$
$$Q_t = X_t \cos \omega t \qquad 7.60$$

b) A low pass filter F is applied on P_t and Q_t so that we get

$$R_t = F(P_t) \qquad 7.61$$
$$S_t = F(Q_t) \qquad 7.62$$

c) Define

$$A_t = 2(R_t^2 + S_t^2)^{1/2} \qquad 7.63$$

$$\tan B_t = -\frac{R_t}{S_t} \qquad 7.64$$

The graphs of A_t and B_t are the amplitude diagram and the phase diagram, respectively.

d) The estimated component of the sequence $\{X\}$ corresponding to the frequency ω is

$$W_t = A_t \cos(\omega t + B_t) \qquad 7.65$$

or similarly

$$W_t = 2(R_t \sin \omega t + S_t \cos \omega t) \qquad 7.66$$

Formula 7.64 is more convenient for the evaluation of the component W_t, whereas formula 7.63 shows the rôles of A_t and B_t.

The choice of an appropriate low pass filter is very important for the whole procedure. Special filters for other frequencies need not be constructed. The transformation given in formula 7.60 enables the use of a low pass filter F acting on the new sequences $\{P_t\}$ and $\{Q_t\}$. In view of the structure of the hydrological sequences, a filter consisting of two moving averages of length k and m has been chosen. The determination of k and m depends on the length of the analysed sequence and on its structure. Information on the structure can be obtained from the periodogram.

The açtual application of the model consists of the following steps:

a) A sequence $X_1, X_2, \dots, X_N$ of real numbers is given.

b) The values

$$X = \frac{1}{N}\sum_{i=1}^{N} X_i \qquad \text{and} \qquad S^2 = \frac{1}{N}\sum_{i=1}^{N} X_i^2 - (\overline{X})^2 \tag{7.67}$$

are calculated. A new sequence $x_1, x_2, \dots, x_N$ is defined by the formula $x_t = X_t - \overline{X}$, $t = 1, 2, \dots, N$. If wanted, the mean X is not substracted, so that $x_t = X_t$. This latter possibility occurs when we are sure that $\{X_t\}$ varies around zero.

c) Frequencies $\omega_1, \dots, \omega_p$ from $(0,\pi)$ are determined by a method based on the periodogram (Anděl, Balek [9]).

d) Natural numbers k and m characterizing the filters are determined; $k + m$ must be an even number.

e) For $\omega = \omega_1$

$$P_t = x_t \sin \omega t \qquad \text{and} \qquad Q_t = x_t \cos \omega t, \; t = 1, 2, \dots, N$$

are calculated.

f) Put

$$C_t = \frac{1}{k}\sum_{i=0}^{k-1} P_{t+1}, \qquad D_t = \frac{1}{k}\sum_{i=0}^{k-1} Q_{t+1} \quad \text{for } t = 1, 2, \dots, N - k + 1$$

$$R_t = \frac{1}{m}\sum_{i=0}^{m-1} C_{t+1}, \qquad S_t = \frac{1}{m}\sum_{i=0}^{m-1} D_{t+1} \quad \text{for } t = 1, 2, \dots, N - k - m + 2$$

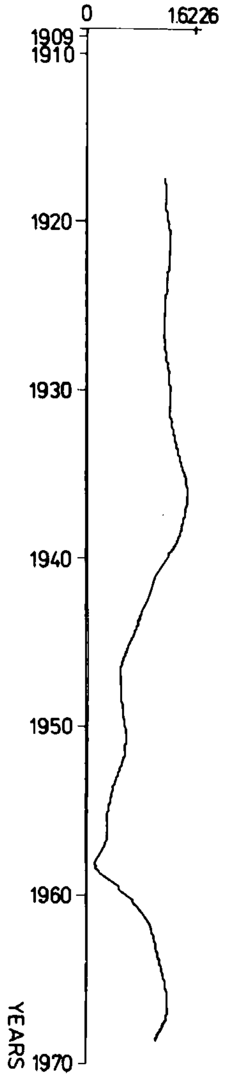
$$\left.\begin{aligned} A_t &= 2(R_t^2 + S_t^2)^{1/2} \\ G_t &= \tan^{-1}\left(\frac{-R_t}{S_t}\right) \\ B_t &= \begin{cases} G_t, & S_t \geq 0 \\ G_t + \pi, & S_t < 0 \end{cases} \end{aligned}\right\} \quad \text{for } t = 1, 2, \dots, N - k - m + 2$$

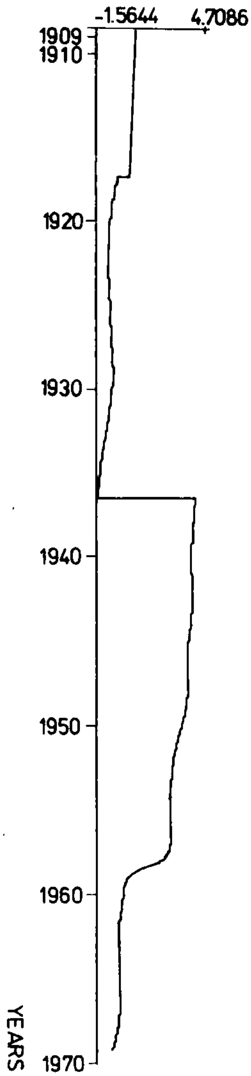
g) The amplitude diagram A_t and the phase diagram B_t are printed. Since A_1 and B_t correspond with time $(k + m)/2$, the time shift is done automatically and for $t < (k + m)/2$ zeros are printed. Similarly, the same number of zeros are printed at the end. We see that $A_t \geq 0$, $-\pi/2 < B_t \leq 3\pi/2$.

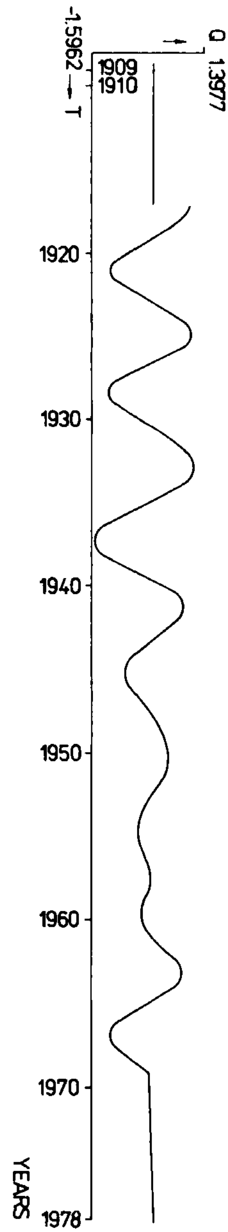
h) The sequence $\{W_t\}$ calculated according to the formula

$$W_t = 2\left[R_t \sin \omega\left(t + \frac{k+m}{2} - 1\right) + S_t \cos \omega t\left(t + \frac{k+m}{2} - 1\right)\right]$$

for $t = 1, 2, \dots, N - k - m + 2$ represents the estimated component in time







7.4 Amplitude diagram at the frequency of 0.0679 for the Sadový spring.

7.5 Phase diagram at the frequency of 0.0679 for the Sadový spring.

7.6 Dumped periodic component at the frequency of 0.0679, Sadový spring.

$(k + m)/2$ and therefore the automatic time shift and the printing of zeros are realised similarly as in the case of A_t and B_t. Then we put $\omega = \omega_2$ and proceed in the same way for all other frequencies $\omega_2, \ldots, \omega_p$.

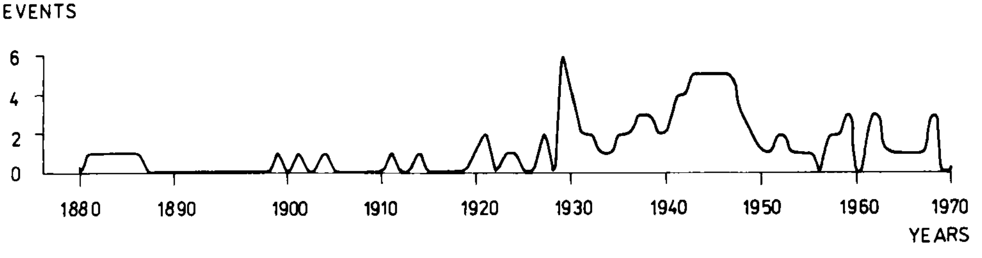
An example of the amplitude and phase diagram is given for the analysis of the periodic components corresponding to a frequency 0.0679, found in the time series of the discharge for the Sadový spring (Fig. 7.4 and 7.5).

Significant disturbances of the amplitude occurred in 1939, 1957 and 1958. The phase was disturbed only in 1957/58; however, the disturbance was very strong.

In Fig. 7.6 is plotted the dumped periodic component for the frequency of 0.0679 corresponding with the above disturbances.

The method was applied in a similar way to the sequences of discharge, water quality, and water temperature of most springs listed in Tab. 7.1.

Fig. 7.7 shows a relationship between the number of disturbances of both types (found in the discharge sequences of the springs) and time. It can be seen that significant disturbances occurred during periods 1929–1932 and 1935–1949 with a peak in 1937–1938 and 1943–1946. Minor disturbances occurred in 1921, 1952, 1957, 1959 and 1978. Several of these disturbances coincided with events associated with mining activity in the vicinity of recharge areas. Such a conclusion may become a solid basis for analysis of the land use impact in the region.



7.7 Occurrence of disturbances in the spring régimes at Karlovy Vary.

It is not recommended, however, to search in advance for events which may have contributed to the occurrence of the disturbances, because our conclusions might become biassed.

7.7 MULTIDIMENSIONAL MARKOV MODEL

Often we need to trace, in the hydrological régimes, the relationships between two or more time series, one of them dependent on the other ones but lagging behind. For instance, we may need to ascertain whether the groundwater level fluctuation is under the impact of a previous pattern of rainfall régime, air pressure and air temperature. For such purpose the so-called Markov multidimensional model has been successfully applied (Anděl, Balek [15]).

It is assumed that p time series are available for the analysis:

$$y_{11}, y_{12}, \dots, y_{1N}$$
$$\dots\dots\dots\dots\dots\dots$$
$$\dots\dots\dots\dots\dots\dots$$
$$y_{p1}, y_{p2}, \dots, y_{pN}$$

We expect that the first time series is influenced by all sequences, the second one by all sequences except the first one, and so on. The last sequence is independent of all the others. Column vectors are formed so that

$$x_1 = (x_{1t}, x_{2t}, \dots, x_{pt}); \qquad t = 1, 2, \dots, N$$

Symbol $'$ means transposition. Provided $A_0, A_1, \dots, A_n$ are matrices of order $p \times p$ and A_0 is regular, we say that the time series x_{it} is controlled by a p-dimensional autoregressive model if

$$\sum_{j=0}^{n} A_j X_{t-j} = \varepsilon_t \tag{7.69}$$

Here ε_t are p-dimensional noncorrelated random vectors with mean values equal to zero and unit variance matrices. If the matrix $A_n \neq 0$ we say that the autoregressive model is of the order n.

The model also can be written as

$$x_t = \sum_{j=1}^{n} U_j x_{t-j} + \xi_t , \tag{7.70}$$

where

$$\begin{aligned} U_j &= -A_0^{-1} A_j , \qquad j = 1, 2, \dots, 4, \\ \xi &= -A_0^{-1} \varepsilon_t \end{aligned} \tag{7.71}$$

The significance of each lag considered in the model is tested by the χ^2 test at 5% level. If the calculated critical value for an autoregression of a certain order is lower than the critical one, the hypothesis on the significance of the autoregression of higher order (or, in other words, the impacts of phenomena more distant in time) is rejected. Beside the χ^2 test, additional information is supplied by a comparison of the variance with residual variance. Sometimes the χ^2 test provides information that an autoregression of certain order is significant, however, from a practical viewpoint the residual variance is not decreasing. Then a simple model based on a smaller number or lags is accepted.

An example of the model application is given for the analysis of the relationship between discharge régimes of the springs at Františkovy Lázně (Czechoslovakia), and air pressure, precipitation and air temperature. Two years of daily observation of all phenomena involved was available and the relationship was traced back for thirty lags.

Using the Markov multidimensional model, we obtain the following information:

$\bar{Q}$	Variance	Residual variance		
$\mathrm{l\ s^{-1}}$		1 lag	3 lags	5 lags
33.982	7.296	3.804	3.266	3.265

The χ^2 value for 5 lags and 16 degrees of freedom was estimated as 31.2, while the critical value at the 5% level of significance is 26.3; however, through a comparison of the residual variances we can see that the model based on 3 lags can be considered as sufficient for practical purposes:

$$\begin{aligned}(Q^t - \bar{Q}) = \; & 0.398\,(Q^{t-1} - \bar{Q}) + 0.039\,(T^{t-1} - \bar{T}) + 0.002(A^{t-1} - \bar{A}) + \\ & +0.0248(S^{t-1} - \bar{S}) + 0.209\,(Q^{t-2} - \bar{Q}) - \\ & -0.040\,(T^{t-2} - T) + 0.0017(S^{t-2} - \bar{S}) + 0.216(Q^{t-3} - \bar{Q}) + \\ & +0.347\,(T^{t-3} - \bar{T}) + 0.016\,(S^{t-3} - \bar{S})\end{aligned}$$

Here Q is the discharge, in $\mathrm{l\ s^{-1}}$; T is the air temperature, in °C; A is the air pressure, in Pa; and S is the precipitation, in mm.

A similar analysis based on mean monthly values provided entirely different information. The impact of the air pressure diminished, and the impact of precipitation and air temperature became more pronounced – five lags back. From both results we can conclude that the spring régime is at least partly influenced by the climate.

7.8 SPATIAL VARIATION

Some soil and aquifer properties vary from place to place, both laterally and vertically. For regionalization purposes the boundaries are plotted rather abruptly and the characteristics are determined from average or single measurements within these boundaries. Such an approach takes little account of gradual changes either within a subregion or from subregion to subregion. As stated by Webster and Burgess [16], these changes can be regarded as properties of the aquifer, or soil, like the depth of the water table and soil water potential.

Two basic features can be recognized:

1. The absence of a repeating pattern – the larger the area or the more intensive the sampling, the more complex is the variation.

2. The point-to-point variation in a sample reflects the real variation in the soil.

Therefore, what is called the regionalized variable theory has been developed, based on the assumption that the values of an aquifer soil property at places close together are likely to be similar, whereas those at places far apart are not.

Let x denote a place in a chosen dimension, and the model of spatial variation can be expressed as

$$z(x) = m(x) + \varepsilon(x)$$

where $z(x)$ is the value of some physical feature Z at x; $m(x)$ is a deterministic function of Z at x; and $\varepsilon(x)$ is a residual of the function. The latter should not be considered a random error but rather local component of variation, and spatially dependent.

The expected value $E[2(x)]$ of a soil property within a subregion is effectively constant, because the sampling variation is often so large that it may be difficult to distinguish an expected change from the noise. For practical purposes, the property is taken as stationary in the mean:

$$E[z(x)] = \mu \qquad 7.72$$

For two places x and $x + h$ separated by vector h

$$E[z(x) - 2(x + h)] = 0 \qquad 7.73$$

It can be expected that the variance of the difference is finite and constant throughout the locality, so that

$$\text{var}\,[z(x) - z(x + h)] = E\{[z(x) - z(x + h)]^2\} = 2\gamma(h) \qquad 7.74$$

The stationary mean and variance of differences define the intrinsic hypothesis of regionalized variable theory, and thus the spatial variation can be described and predicted in a simple way.

The quality $\gamma(h)$ which is half the variance of difference between values at places separated by h, is the semi-variance (a dissimilarity of two places). It can be estimated from data measurements as

$$\hat{\gamma}(h) = \frac{1}{2m} \sum_{i=1}^{m} [z(x_i) - z(x_i + h)]^2 \qquad 7.75$$

where m are pairs of observations at lag h in space.

A graph of $\hat{\gamma}(h)$ against h is a sample semi-variogram. The procedure in which the semi-variogram is used for interpolation is known as kriging (in honour of D. C. Krige – see Journal and Huijbregts [17]). Kriging is defined as a local estimation in which each estimate is a weighted average of a observed values. The estimates refer to points as, for instance, to the volume of soil the same size and shape as those on which the measurements were taken. The estimated value of property Z at place x_0 is defined as

$$\hat{z}(x_0) = \sum_{i=1}^{n} \lambda_i \, z(x_i)\,, \qquad 7.76$$

The weights λ_i are chosen so that the estimate $\hat{z}(x_0)$ is unbiased and the estimation variance lower than for any other linear combination of the observed values. Thus the weights add up to 1 and the minimum variance

$$\sigma_E^2 = \sum_{j=1}^{n} \lambda_j \gamma(x_j, x_0) + \psi \qquad 7.77$$

is obtained when

$$\sum_{j=1}^{n} \lambda_j \gamma(x_i, x_j + \psi = \gamma(x_i, x_0) \quad \text{for all } i \qquad 7.78$$

The quantity $\gamma(x_i, x_j)$ is the semi-variance of Z between the sampling points x_i and x_j while (x_i, x_0) is the semi-variance between the sampling point x_i and the point x_0, both obtained from the semi-variogram. The quantity ψ is a Lagrange multiplier associated with minimization.

Similarly, for block estimates the minimum variance is

$$\sigma_B^2 = \sum_{j=1}^{n} \lambda_i \bar{\gamma}(x_j, x_B) + \psi_B - \bar{\gamma}(x_B, x_B) \qquad 7.79$$

obtained when

$$\sum_{j=1}^{n} \lambda_j \gamma(x_i, x_j) + \psi_B = \bar{\gamma}(x_i, x_B) \quad \text{for all } i \qquad 7.80$$

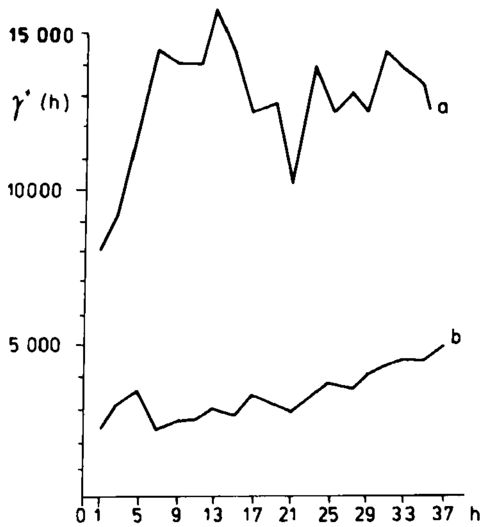
Here the quantity $\bar{\gamma}(x_i, x_B)$ is the average semi-variance between the sampling point x_i and all points within B, and $\bar{\gamma}(x_B, x_B)$ is the variance within B.

Solving the above sets of equations we obtain weights which can be used to find the estimated value for a given place.

As an example can be given the application of the method on the analysis of the groundwater level amplitudes in Czech sandstone formations. The amplitude may be at the first sight considered as purely hydraulic phenomenon, however, it provides significant information on the structure of the upper layers of the lithosphere. Valuable conclusions can be obtained on the variability of the recharge process and depletion of the resources (Balek, Sobíšek [22]).

As $z(x_i)$ was taken the difference between maximum and minimum of observed groundwater level during a representative period 1974–1976. Considering the non-uniform distribution of the boreholes in a very extensive area h was represented in the interval $\langle h - 1, h + 1 \rangle$ km and the distance involved was 2 km. Two separate semivariograms were obtained for eastern and western part of the region, regarding a fact that the groundwater circulation has a different pattern in each part. Number of pairs was in each case at least 176. As it can be traced in each graph (Fig. 7.8) the variability of the examined variable is greater in western part of the region (graph a). According to the rise of this semivariogram up to the distance of seven kilometres it can be concluded that the amplitudes are mutually correlated at the distance of no more than five kilometres.

Data do not contravene the presumption that the fluctuation of the groundwater level is rather spatially independent, otherwise the semivariograms would increase explosively with increasing *h*.



7.8 Semivariograms of the groundwater level amplitude for western (a) and eastern (b) part of Czech Turonian formations.

Spatial variability analysis of the properties of surface can also contribute to the tracing of the recharge areas and preferred pathways. For instance, an intensive groundwater recharge may occur where the soil profile is rather shallow or non-existent. Denuded rocks with exposed fissures and fractures on the surface can favour intensive recharge. Natural depressions on the surface are also indicators of the preferred pathways.

An important rôle can be played by the slope of the surface, and of the boundaries between soil layers and weathered rock. In fact, this factor is considered significant in the theory of infiltration barriers (Rancon [20]). Andersen and Madsen [21] attempted to establish such a barrier artificially as a protection against the water percolating into waste deposits. In principle, they assumed that there was no vertical flow from an almost saturated fine-textured material to an unsaturated coarse-textured material, until a specific saturation level dependent on the capillary rise had been formed. If the boundaries between two layers are sufficiently steep, the infiltrating water is drained laterally through the upper fine-textured layer. The authors believe that any slope steeper than 1% can be sufficient to form the capillary barrier, provided the top layer has a high hydraulic conductivity and high capillary rise.

7.9 FUTURE DEVELOPMENT

The models presented can only illustrate some of the possible trends in the development of stochastic models and their application in hydrogeology. It is foreseen that future development will be concerned with combining various types of theories on one hand, and using the models to imitate the decision making of their users on the other.

An example of the first trend is the stochastic system model of groundwater level fluctuation, conducted by Adamowski et al. [18], based on a complex input-output view of all processes involved. The second trend can be exemplified by the ARM model. ARM is an acronym for "Automata with Random Mutations" (Anděl et al. [19]). In contrast to the previous models based on a fixed logistic algorithm, the ARM imitates the intellectual approach of the model maker. The model is constructed so that prediction of a certain phenomenon one step ahead is based on the model's own continuously increasing and improving experience. In principle, those results achieved by the model which have been found successful for the initial part of a given time series are further utilised, while the unsuccessful ones are eliminated by the model itself. Thus only the experience of the model within a given time series serves for the further improvement of results. The experience is evaluated using special criteria.

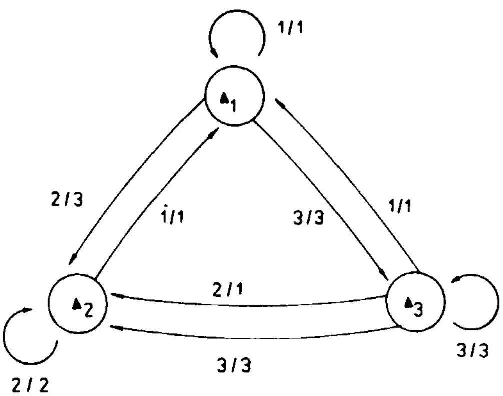
Input information is entered in the form of a time series consisting of certain symbols defined as the input alphabet. For instance, the régime of the groundwater level to be simulated by ARM can be simplified so that each discharge falls into one of the categories marked, for instance, 1; 2; 3; and indicating low, mean and high groundwater level. The model assumes a series of states from the set of states A_1, A_2 and A_3 and produces an output from the set of categories 1, 2, 3; at discrete time intervals. At any time the next output and next state are determined by the current state and output, according to a set of rules such as "If in state A_1 with output 2 go to new state A_2 with new output 3". The complete set of rules may be expressed as:

$$\begin{array}{ccccccccccc} A_1 & A_1 & A_1 & & A_2 & A_2 & A_2 & & A_3 & A_3 & A_3 \\ 1 & 2 & 3 & & 1 & 2 & 3 & & 1 & 2 & 3 \\ 1 & 3 & 3 & & 1 & 2 & 3 & & 1 & 1 & 3 \\ A_1 & A_2 & A_3 & & A_1 & A_2 & A_3 & & A_1 & A_2 & A_3 \end{array} \tag{7.81}$$

The first row indicates the current state of the automaton and the second row the current output. The third and fourth row define the new output and the new state of the automaton. Operation of the automaton is shown schematically in Fig. 7.9. For example, if the phenomenon is in current state A_1 with current output 2, new state A_2 with output 3 will follow.

It is feasible to construct the automaton so that it reflects the observed behaviour of the studied phenomenon. This process is called adaptation of the automaton to a given environment. New automata are formed with minor changes when compared with the previous ones and the set of outputs is compared with the time series. Then the automaton producing the best agreement with actual values is chosen for further analysis.

The changes of the automata are provided at random, and the process is literally analogous to the breeding of plants or animals when only the best offspring is used for further breeding.



7.9 A graphical interpretation of the automaton.

Let us take the time series $x_1, x_2, \ldots, x_n$ composed of the symbols of the input alphabet. Let us take a natural number h less than n. It is assumed that in the course of simulation of the time series, an automaton with m states has been obtained. Such an automaton is called parental. Changes (mutations) which can be further performed on it are of the following types:

1. change of the initial state,
2. change of some of the output symbols in a certain column of the automaton,
3. change of the state to which the automaton for a certain column can be transferred,
4. the addition of another state,
5. the omission of one state.

For computer analysis it is necessary to specify in advance how many mutations are to be performed, how each mutation is to be set up, and how a new automaton will be applied. A statistical background of the mutations is analysed by Anděl et al. [19]. The performance of ARM is always tested from the very beginning. The measure of the predictive ability of each automaton created, including the parental one, is obtained by comparing the results with the analysed sequence. When programming the algorithm of ARM, Simula has been found to be the most effective language.

An example of the application can be given on the sequence of 832 mean monthly discharges of the Svobody spring, observed from August 1908 till December 1977 at Karlovy Vary. As in all other cases the initial automaton was set up in accordance with 7.81. However, the final automaton has two states only:

A_1	A_1	A_1	A_2	A_2	A_2
1	2	3	1	2	3
1	2	3	1	3	3
A_2	A_1	A_1	A_1	A_2	A_2

Similarly, for the sequence of mean monthly temperature of the Svobody spring, the following automaton was found:

A_1	A_1	A_1	A_2	A_2	A_2
1	2	3	1	2	3
1	3	3	1	3	3
A_2	A_2	A_2	A_2	A_2	A_1

The difference between both automata may indicate a certain independence of the temperature and discharge régimes.

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8 GROUNDWATER ASSESSMENT AND MANAGEMENT

8.1 GENERAL

A knowledge of how the groundwater régime behaves under given environmental conditions is essential for further groundwater resources development. Any groundwater surveys, data collection, analysis, expeditional measurements, simulation methods and other means of the system approach should always be performed with proper respect to the practical objectives which are to be achieved in the given aquatic ecosystem. Thus an assessment of groundwater availability extends far beyond the stage of data collection. It also involves an analysis of groundwater flow and its interaction with the stream network, an assessment of the hydrological balance, and of the time of residence. However, it also includes forecasting the future use of water resources based on various exploitation alternatives. As well as technical and hydrological analysis, the economical and social aspects must be taken into account. The evaluation of future demands should be based upon the results of a comprehensive survey among the users. Such an approach will vary from region to region, and thus the criteria involved cannot be blueprinted for a great number of countries. For instance, a sensitive approach which recognizes the demand of local customers is an important part of the survey in many developing countries (Balek [1]). Also, approaches can be expected to differ in humid and arid regions. In a humid region flood control may have priority in water planning, and this may lead to the conclusion that there is always abundant groundwater or because surface waters are available, the groundwater is considered to be a less significant source. However, in these regions groundwater is often the only source of supply during a long or short dry period, or when surface water becomes heavily polluted and its treatment expensive. Groundwater is always a reasonable source of safe water supply for rural areas and for domestic use. Groundwater becomes a more significant source in semi-arid and arid lands where it is often the only water resource available. In such areas even a costly, large-scale survey which includes aerial photography, field geological survey, groundwater data collection and evaluation, geophysical investigation, and isotope studies, is usually profitable.

When trying to establish the critical amount which can be extracted from the aquifer as part of renewable resources, the use of existing observations is essential. Only after they have been evaluated can newly established observational programmes be planned together with other survey methods, mainly to fill the gaps in existing data and in the identification results. At present such work is directed towards

Tab. 8.1 Groundwater level, mean annual discharge, and mean annual baseflow in the Svitava river basin, northern Czechoslovakia.
VH – very high, H – high, O – normal, L – low, VL – very low.

Year	Mean gw. level m	Régime	River discharge $m^3 s^{-1}$	Régime	Baseflow $m^3 s^{-1}$
1931	16.86	VH	2.834	VH	1.916
1932	19.47	H	1.935	O	1.635
1933	21.68	L	1.417	VL	1.397
1934	23.59	VL	1.251	VL	1.192
1935	22.34	VL	1.583n	L	1.326
1936	21.13	L	1.671	L	1.456
1937	19.96	O	2.258	H	1.582
1938	18.38	H	2.844	VH	1.752
1939	16.61	VH	2.844	VH	1.953
1940	17.29	H	2.532	H	1.870
1941	17.20	VH	3.441	VH	1.879
1942	17.40	H	2.610	H	1.858
1943	20.85	L	1.486	L	1.481
1944	20.66	O	1.838	L	1.507
1945	19.41	H	2.268	H	1.641
1946	19.23	H	2.307	H	1.661
1947	19.47	O	2.111	H	1.635
1948	18.97	H	2.219	H	1.689
1949	20.83	L	1.535	L	1.489
1950	21.60	L	1.496	L	1.406
1951	21.00	L	1.896	L	1.470
1952	20.93	L	1.730	L	1.478
1953	19.62	O	2.160	H	1.619
1954	23.16	VL	1.662	L	1.238
1955	20.90	L	2.209	H	1.481
1956	21.05	L	1.896	L	1.465
1957	18.80	H	2.228	H	1.707
1958	19.52	O	2.317	H	1.630
1959	19.20	H	1.906	L	1.664
1960	20.67	L	2.023	O	1.506

establishing data banks. Haman [2] listed the aims which can be achieved through data banks as follows:

a) to provide a progressive build-up of knowledge of groundwater conditions in the area,
b) to provide a reliable background for water resources master planning and the selection of the most promising areas for groundwater resources development,
c) to optimize the number of wells, drilling depth, and well construction,

d) to protect groundwater from pollution,
e) to control groundwater quality,
f) to improve the management of groundwater resources.

In the sections which follow, various aspects of groundwater management and exploitation are discussed, which are based upon the conclusions reached in preceding chapters.

8.2 PROBABILITY CONCEPT IN SAFE YIELD ANALYSIS

When the maximum safe yield has been determined for an approximately representative period of time, the question to be answered is, to what extent can the groundwater resources be exploited during the dry years? If there is a substantial amount of dead storage in the area, the sustained yield can be supplied from the dead storage for some time, and later, during the wet years, the groundwater storage can be replenished from excess groundwater recharge. When only live storage is available, then during the dry years the actual yield will be quite low, while in wet years it will exceed the safe yield. This problem is closely related to the probability of occurrence of low river discharges, because during prolonged dry period a great part of the river discharge, if not all, is supplied from groundwater resources.

While it is relatively easy to determine the empirical or theoretical probability of occurrence of river discharges, there are more problems involved in assessing the probability of occurrence of baseflow.

Some information can be obtained from the long-term fluctuations of groundwater level. In Tab. 8.1 the mean annual groundwater level observed during the representative period 1931 – 1960 in the Banín borehole (in northern Czechoslovakia) is shown. Tab. 8.2 gives the calculated empirical probability of occurrence of the mean groundwater level, while Tab. 8.3 gives the theoretical probability of occurrence based on Pearson's type III distribution.

Each of the mean levels in Tab. 8.1 can be characterised in accordance with Tab. 8.4 (Netopil [3]). Also given in Tab. 8.1 is the sequence of mean annual discharges observed in a nearby river, and Tab. 8.3 includes the calculated theoretical probability of occurrence of mean annual discharges of the same river.

The maximum safe yield for the period 1974 – 176 was estimated as $Q_g = 1.526\,\mathrm{m^3\,s^{-1}}$ for the recharge area $A = 193.64\,\mathrm{km^2}$. The recession coefficient $K = 0.999\,98$. The mean groundwater table depth during that short period was 20.48 m, which according to Tab. 8.2 and 8.4 can be considered as the normal level.

Thus the mean active groundwater storage in the area is

$$G = \frac{-Q_g \cdot 3.6}{A \ln K} = 1418.5\,\mathrm{mm}$$

The groundwater zone capacity is

$$SG = (H \times N) + G = 3466.5\,\text{mm},$$

where H is the mean groundwater level, in mm; and N is the noncapillary porosity ($N = 0.10$).

Tab. 8.2 The empirical probability of occurrence of the groundwater level and baseflow

m	Gw level in ascending order, metres	$p = \frac{m}{n+1}$	Baseflow descending order $m^3 s^{-1}$
1	16.61	3.2	1.943
2	16.86	6.5	1.916
3	17.20	9.7	1.879
4	17.29	12.9	1.870
5	17.40	16.1	1.858
6	18.38	19.4	1.752
7	18.80	22.6	1.707
8	18.97	25.8	1.689
9	19.20	29.0	1.664
10	19.23	32.3	1.631
11	19.41	35.5	1.641
12	19.47	38.7	1.635
13	19.47	41.9	1.635
14	19.52	45.2	1.630
15	19.62	48.4	1.619
16	19.96	51.6	1.582
17	20.66	54.8	1.507
18	20.67	58.1	1.506
19	20.83	61.3	1.489
20	20.85	65.0	1.481
21	20.90	67.7	1.481
22	20.93	71.0	1.478
23	21.00	74.2	1.470
24	21.05	77.4	1.465
25	21.13	80.6	1.456
26	21.60	83.9	1.406
27	21.68	87.1	1.397
28	22.34	90.3	1.326
29	23.16	93.5	1.238
30	23.59	96.8	1.192

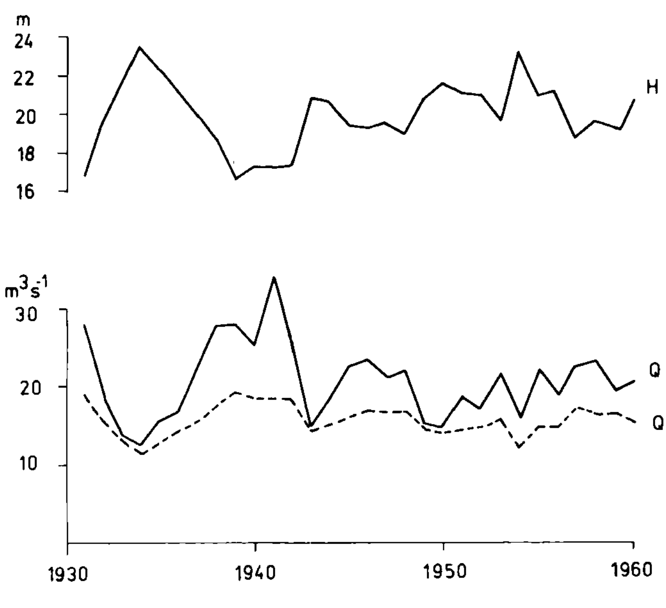
For the groundwater levels of certain probability of occurrence (Tab. 8.2 and 8.3) the safe yields of corresponding probability can be determined by substituting the appropriate mean groundwater levels. The results are shown in Tables 8.1, 8.2,

Tab. 8.3 The theoretical probability of occurrence of mean groundwater level, river discharge, and baseflow in the Svitava basin, northern Czechoslovakia

Probability of occurrence, %	1	5	10	20	50	70	90	95	99	Cv
Groundwater level, m	16.3	17.0	17.6	18.1	19.8	20.9	22.2	23.0	24.4	0.0889
River discharge, $m^3 s^{-1}$	3.49	2.91	2.64	2.35	1.81	1.58	1.36	1.27	1.12	0.242
Baseflow, $m^3 s^{-1}$	1.98	1.90	1.83	1.78	1.60	1.49	1.34	1.25	1.10	0.131

and 8.3. By comparing the mean annual discharge and mean baseflow (Tab. 8.3) we can see that if the maximum safe yield is exploited the stream will become almost intermittent in 30 years out of a hundred.

The graphical interpretation of the results is in Fig. 8.1. Comparing the régimes of the surface and groundwater flow in the river (Tab. 8.1) we can see that both are relatively independent particularly in wet years, when the stream is influenced by floods and an increased subsurface flow.



8.1 Groundwater level below the surface H, mean annual discharge Q_i and baseflow Q_g in the Svitava river during the representative period 1931 – 1960.

A different pattern emerges when the safe yield is compared with the probability of occurrence of minimum daily discharges. For a given basin it follows that the minimum daily discharge is likely to be equal to, or less than, the values given below:

Once in	10	20	50	100	years
$Q_{d_{min}}$	.594	.539	.493	.456	$m^3 s^{-1}$

We should bear in mind that the occurrence of minima is under the strong influence of evapotranspiration, particularly in the vinicity of a stream network where the incoming water is close to the surface (Garczynski [4]). In many places, an uncontrolled pumping of water from the streams during dry periods complicates the situation. Thus the régime of the occurrence of minima cannot be related directly to the occurrence of safe yield unless the above phenomena are taken into account. As in the example given above, the minimum baseflow likely to occur once in 100 years ($H = 25.4$ m) should be $1.10\ m^3\ s^{-1}$.

Tab. 8.4 Characteristics of the groundwater level régime (after Netopil)

Probability of occurrence, %	Régime	Symbol
0 – 10	very high	VH
11 – 40	high	H
41 – 60	normal	O
61 – 90	low	L
>91	very low	VL

Therefore, any methods of safe yield assessment which are based on the analysis of daily minimum discharge may lead to misleading conclusions.

When long-term groundwater level records are absent, the occurrence of safe yield can be deduced from precipitation records. Tab. 8.5 gives the calculated

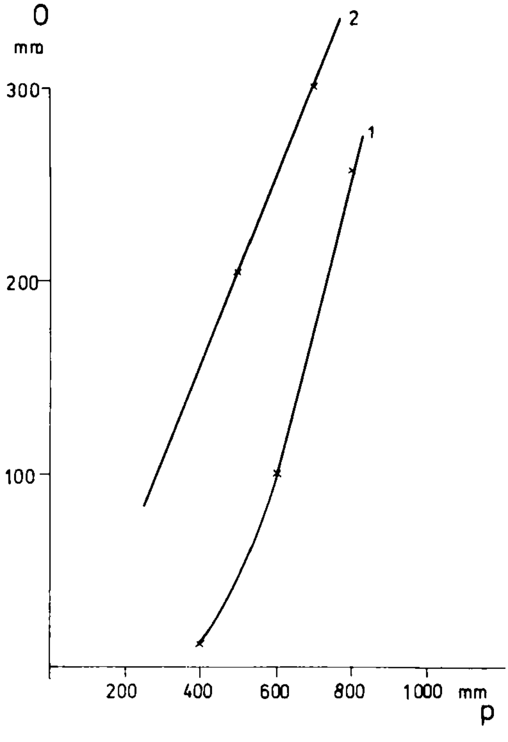
Tab. 8.5 The theoretical probability of occurrence of mean annual precipitation baseflow for a stabilised and an unstable groundwater régime in the Jizera basin, northern Czechoslovakia (See also Fig. 8.2).

Probability of occurrence, %	1	5	30	50	70	95	99
Stabilised régime:							
Annual precipitation, mm	1009	897	757	694	631	512	456
Baseflow, $l\ s^{-1}$	2481	2160	1742	1569	1364	1011	838
Unstable régime:							
Annual precipitation, mm	844	730	616	564	513	416	371
Baseflow, $l\ s^{-1}$	1810	1226	704	490	342	101	25

Credibility limits at 95% level of significance for the probability of occurrence equal to 50%:

Precipitation mm	Baseflow $l\ s^{-1}$	Régime
735 – 563	1676 – 1440	stabilised
602 – 526	600 – 376	unstable

theoretical probability of occurrence of mean annual rainfall, and in Fig. 8.2 is plotted the rainfall-baseflow relationship as determined by a conceptual model for an annual precipitation of varying probability of occurrence. Thus the corresponding baseflow has a corresponding probability of occurrence, however, because the properties of the aquifers must be taken into account, the probability will differ from one area to another.



8.2 Mean annual rainfall-baseflow relationship produced by the deterministic model for shallow (1) and deep aquifer (2).

According to the sampling theory we should bear in mind that a mean annual precipitation based on a short record should not be used as an absolute value, and that credibility limits should be introduced. In the case given as an example, the credibility limits for mean annual precipitation are 755 – 653 mm, and the corresponding safe yield is determined within the limits of $1.676-1.440\,m^3\,s^{-1}$ for the stabilised régime.

8.3 EFFECTS OF LAND USE UPON THE GROUNDWATER RÉGIME

Any problems of land use impact are closely related to the problem of how man's activities influence the whole hydrological cycle. The natural groundwater régime is closely associated with climatic conditions and with the nature of the aquatic ecosystem. As the population increases industrialisation and urbanisation occupy more and more land surface; at the same time more arable land is needed for food production. Thus vast areas are deforested which, particularly in tropical regions, has a tremendous impact on hydrological régimes. Reclamation of waste land also contributes significantly to rapid changes in the water cycle. The construction of reservoirs, canalisation of rivers and initiation of drainage and irrigation schemes produce secondary effects (see subsection 5.6.5) which must be evaluated as an integral part of any groundwater resources appraisal.

It is clear that most of the impacts are mutually related, and that those of primary importance are accompanied or followed by other impacts which although less significant at first sight, may become of primary importance when their final effects are considered.

Colebrander [5] recognizes three basic types of impact. He considers as primary those impacts which have been brought about intentionally in a deliberate attempt to change the hydrological régime or parts of it. Typical examples are irrigation, dam construction, and river training. He considers as secondary those impacts which are not associated with deliberate attempts to change the hydrological cycle, but which in many ways may contribute to its change. Examples include the construction of canals, embankments, and water recreation facilities.

The third type of impact, inadvertent activities, might be considered as independent of the hydrological régime but in fact they may affect it considerably in an indirect way. Many human activities related to agriculture and forestry management belong here, and an evaluation of their influence on the groundwater régime is one of the most difficult tasks in hydrological analysis.

Many attempts have been made to classify the land use aspects in effective water management. King [6] set up a general land classification intended for wide-scale application. He differentiated land classification, which allocates similar types of land into particular classes, from land-use planning which combines land classi-

Tab. 8.6 Check-list of minimum requirements for land classification, after King

Soil	Topography
Profile	Elevation
Texture	Degree of slope
Structure	Aspect
Chemical reaction	
Content of organic matter	
Content of essential plant nutrients	
Depth	
Land condition	Climate
Erosion	Precipitation
Deposition	Temperature
Depletion	Climatic hazards

fication with economic, legal, social and institutional factors, and is used in decision-making. According to King's approach, such classification is concerned only with the soil (Tab. 8.6).

Eren [7] modified the above approach by including what can be called the socio-economic features of a catchment (Tab. 8.7).

Tab 8.7 Some parameters in a socio-economic survey of a catchment, after Eren.

Population census and description	
total population	rate of growth
sex ratios	migration patterns
age structure	
Behavioural and social characteristics	
family pattern	reaction to innovation
family size	educational level
traditions	religious sects
taboos	work ethics
details on communal administration	health schemes
farm/village organisations	forestry organisations
Economic and marketing factors	
land tenure patterns	marketing arrangements
patterns of cultivation	crop surveys
farming practices	yields of crops
shifting cultivation	labour – settled, migrated
transport system	local/export use
forest inventories/descriptions	range and grazing surveys
industry	
forest fire hazards/protection	

Even if we have all information required for such classification we are still at the initial stage of land use analysis because we must also know what will happen after the present state of affairs is changed. Thus an impact analysis becomes a rather complicated part of the land use studies. Among the many methodologies should be mentioned what is known as the Leopold Matrix (Leopold et al. [8]) and the Batelle Environmental Evaluation System (Dee et al. [9]).

The Leopold matrix method lists the following actions across the top on the horizontal axis (Tab. 8.8):

Tab. 8.8 List of environmental actions and characteristics in the Leopold Matrix, after Leopold

The actions listed across the top:

Type of action	Example
A. Modification of régime	habitat modification
B. Land transformation	airports, roads
C. Resources system extraction	timber harvest, fishing

Tab. 8.8 cont.

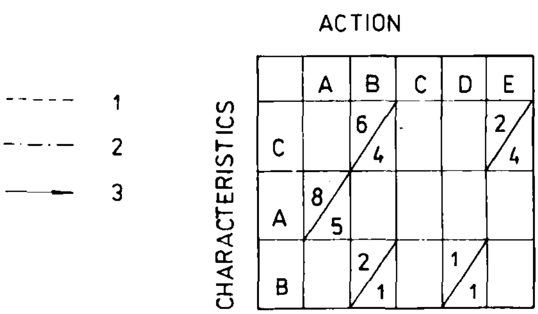
The actions listed across the top:

Type of action	Example
D. Processing	energy generation
E. Land alteration	strip mining, erosion control
F. Resource renewal	groundwater recharge
G. Changes in traffic	automobiles, railways
H. Waste emplacement and treatment	effluent discharge
I. Chemical treatment	fertilisation and pesticides
J. Accidents	oil spills
Environmental characteristics on the vertical axis:	
A. Physical and chemical characteristics	air or water quality
B. Biological conditions	flora and fauna
C. Cultural factors	aesthetics, recreation
D. Ecological relationships	eutrophication, food chains

Along the vertical axis are listed environmental characteristics, or conditions which may be affected by the actions listed above (Tab. 8.8).

Obviously, for a direct application in groundwater resources management the actions and characteristics involved must be chosen according to the outputs requested.

In the matrix (Fig. 8.3) two values are inserted which show on a scale of 1 to 10, the magnitude and the importance of the possible impact. A verbal explanation of the figures should accompany each matrix.



8.3 The Leopold matrix.

The Batelle Environmental Evaluation System is based on a quantitative and objective comparison of the impacts observed, or predicted, for various parts of the total environment. In order to compare the environmental conditions with and without a given activity, it is essential that the impact is measured in commensurate units.

The evaluation system is hierarchial in nature, ranging from a general level of

Tab. 8.9. Battelle environmental evaluation system (Dee et al.)

Environmental impact

Ecology 240

Species and Populations
Terrestrial
(14) Browsers and grazers
(14) Crops
(14) Natural vegetation
(14) Post species
(14) Upland game birds
Aquatic
(14) Commercial fisheries
(14) Natural vegetation
(14) Pest species
(14) Sport fish
(14) Waterfowl
[140]

Habitats and Communities
Terrestrial
(12) Food index
(12) Land use
(12) Rare and endangered species
(14) Species diversity
Aquatic
(12) Food index
(12) Rare and endangered species
(12) River characteristics
(14) Species diversity
[100]

Ecosystems
Descriptive only

Environmental Polution 402

Water Pollution
(20) Basin hydrologic loss
(25) BOD
(31) Dissolved oxygen
(18) Fecal coliforms
(22) Inorganic carbon
(25) Inorganic nitrogen
(28) Inorganic phosphate
(16) Pesticides
(18) pH
(28) Stream flow variation
(28) Temperature
(25) Total dissolved solids
(14) Toxic substances
(20) Turbidity
[318]

Air Pollution
(5) Carbon monoxide
(5) Hydrocarbons
(10) Nitrogen oxides
(12) Particulate matter
(5) Photochemical oxidante
(10) Sulphur oxides
(5) Other
[52]

Land Pollution
(14) Land use
(14) Soil erosion
[28]

Noise Pollution
(4) Noise
[4]

Aesthetics 153

Land
(6) Geological surface material
(16) Relief and topographic character
(10) Width and alignment
[32]

Air
(3) Odour and visual
(2) Sounds
[5]

Water
(10) Appearance of water
(16) Land and water interface
(6) Odour and floating materials
(10) Water surface area
(10) Wooded and geologic shoreline
[52]

Biota
(5) Animals – domestic
(5) Animals – wild
(9) Diversity of vegetation types
(5) Variety within vegetation types
[24]

Man-made Objects
(10) Man-made objects
[10]

Composition
(15) Composite effect
(15) Unique composition
[30]

Human Interest 205

Educational Scientific Packages
(13) Archeological
(13) Ecological
(11) Geological
(11) Hydrological
[48]

Historical Packages
(11) Architecture and styles
(11) Events
(11) Persons
(11) Religions and cultures
(11) Adventure value
[55]

Cultures
(14) Indians
(7) Other ethnic groups
(7) Religious groups
[28]

Mood Atmospheres
(11) Awe-inspiration
(11) Isolation/solitude
(4) Mystery
(11) "Oneness" with nature
[37]

Life Patterns
(13) Employment opportunities
(13) Housing
(13) Social interactions
[37]

Legend ( ) Parameter Importance Units
[ ] Total

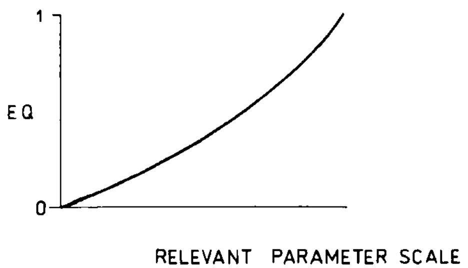
four environmental categories to the detailed level of environmental indicators (Tab. 8.9).

The procedure consists of three main steps:

1. All parameters are transformed into their corresponding level of environmental quality – EQ which ranges from 0 to 1 (Fig. 8.4).

2. The relative importance of parameters involved is expressed in parameter importance units – PIU (Tab. 8.9). The units provide the necessary uniformity between the projects, however, they do place a somewhat empirical constraint on the method as a whole.

3. The environmental impact unit – EIU of action on a given parameter is obtained as $EIU = EQ \times PIU$.



8.4 An example of the value function which is used in the Batelle evaluation system.

Then the total EIU for each major category can be compared for different project alternatives, or for "with" and "without" a given project.

Neither method was developed specifically for groundwater resources assessment under changing land use/environmental impacts, and therefore they may serve only as guidance when a particular groundwater development project is to be evaluated with respect to changing environmental conditions.

Sometimes the assignment of environmental impact units may be difficult. If we intend to study the impact of the groundwater régime in a climatologically and morphologically uniform region, underlined by a rather monotonous geological structure, it is convenient to establish a network of representative catchments in which different types of land use are dominant, and to compare the water balance

Tab. 8.10 Basic features of the representative catchments

Catchment	Drainage area, km^2	Forest %	Field %	Meadow %	Urbanisation %
Hartvikov	0.984	100.0	–	–	–
Pojbuky	2.039	1.5	37.0	40.5	21.0
Vočadlo	0.586	3.0	86.0	11.0	–
Březina	152.690	35.0	32.0	28.0	5.0

equations, particularly their groundwater components. One example of this approach is the study undertaken in the Czech-Moravian Highlands (Balek, Skořepa [10]).

In this region the water balance of forested catchment of natural field, rural and urban catchment, and of mixed basin typical for the region were analysed. Tab. 8.10. shows the basic features of the catchments selected.

Hartvíkov is a forested catchment with dominant coniferous trees, Pojbuky is an agricultural catchment with a considerable part of it covered by rural/urban structures, Vočadlo is a catchment where non-drained fields dominate, and Březina is a basin with mixed land use.

Tab. 8.11 Water balance of the representative catchments

Catchment	Rainfall mm	Surface mm	Subsurface runoff mm	Groundwater mm	ET mm
Hartvíkov	781	10.9	29.7	67.0	673.4
Pojbuky	746	134.0	161.4	101.4	349.2
Vočadlo	736	39.3	75.6	55.6	565.5
Březina	693	115.2	99.8	98.7	379.3

In the course of three-years of observation the following water balance equations were obtained (Tab. 8.11). The values of the characteristic distribution of monthly yield are given in Tab. 8.12. In Tab. 8.13 are the daily discharges likely to be in excess

Tab. 8.12 Monthly yield, in $l\ s^{-1}\ km^{-2}$, of the representative catchments

Catchment	11	12	1	2	3	4	5	6	7	8	9	10
Hartvíkov	3.6	1.0	2.9	3.6	4.7	3.8	3.1	2.6	2.2	4.4	3.8	1.1
Pojbuky	10.7	10.0	19.6	21.3	20.7	13.1	10.2	6.5	3.8	14.4	11.3	9.7
Vočadlo	3.2	3.2	7.0	3.9	11.1	5.1	4.4	3.4	2.0	7.0	6.0	3.8
Březina	5.5	5.6	7.0	23.8	15.7	6.2	5.1	4.0	3.0	15.7	9.3	5.5

in a normal year. The values are given as a percentage of mean annual discharge.

By comparing the results we can see a high stabilising effect of the forest. This, however, is due to the great water loss (due to evapotranspiration). A different extreme can be seen in the régime of the urbanised catchment Pojbuky, in which a very high total runoff and the highest groundwater outflow was observed. This can be explained by a reduced rate of evapotranspiration from paved areas. Concentrated surface runoff from paved areas infiltrates into the soil and contributes to an increase in groundwater and subsurface water resources. Slightly reduced baseflow was observed in the Březina basin, where an additional effect of surface runoff from roads, villages etc. played an important rôle. Causes of the relatively low groundwater outflow from the Vočadlo catchment might be the lower annual precipitation, and the high evapotranspiration during the growing season. In the forested catchment of Hartvíkov, groundwater resources are replenished from snowmelt to a greater extent than the groundwater resources located under the field.

Tab. 8.13 Distribution of mean daily discharge as a percentage of mean annual discharge in the representative catchments

Catchment	Mean daily discharge likely to be exceeded for ... % of a year (% of mean annual discharge)				
	20	40	60	80	90
Hartvíkov	135	96	62	43	41
Pojbuky	182	90	49	24	19
Vočadlo	152	63	61	26	22
Březina	120	60	40	30	21

A higher subsurface flow in the field can be explained by a repeated replenishment of shallow groundwater resources during the growing season, when most of the potential replenishment of shallow groundwater storage in a forest area is captured by the forest interception.

The stabilizing effect of the forest can also be seen in Tab. 8.13, by comparing the duration curves for Hartvíkov and Pojbuky. Here, however, the differences in mean annual discharge and corresponding yield must be taken into account.

Many regions today are faced with extensive and rapid changes which in turn play a significant rôle in the formation of groundwater resources. Due to improper agricultural management a substantial amount of soil is being eroded year by year, and conditions for the optimum groundwater recharge have deteriorated. Vast forest areas are endangered, and large-scale timber felling may cause serious changes of climate; margins of deserts are threatened by desertification and pastoral lands suffer from overgrazing.

These and many other impacts should be systematically monitored and analysed, as a basis for the better management of groundwater resources through optimized land use.

8.4 PROTECTIVE MEASURES IN REGIONS WITH STEADILY DECLINING GROUNDWATER LEVEL

Reports from many parts of the world indicate that non-renewable groundwater resources are being heavily consumed. The differentiation between renewable and non-renewable resources should become an important part of groundwater resources management projects. At first the problem has to be tackled regionally, this being based on complex water balance calculations which take into account the long-term means of annual precipitation, evapotranspiration, surface runoff, and the volume of groundwater which is currently being exploited. In some regions where the ratio between dead and live storage is high, and where the groundwater is exploited through a storage, it may take a long time before the impact of exploitation on the active storage and surface stream network becomes pronounced. By then it is usually too late to implement protective measures, because the population has become accustomed to using groundwater resources in excess of the safe yield.

Dijon [11] listed the following protective measures which can lead to increased water availability, particularly in semi-arid and arid regions:

1. reducing direct evapotranspiration by lowering the groundwater table,
2. changing cultivation methods which, when properly applied, may reduce transpiration from the saturated zone,
3. developing facilities for the treatment of polluted water which otherwise would not be usable,
4. artificial recharge,
5. optimizing water extraction, particularly through the proper choice of either open hand-dug wells or drilled boreholes. Although construction of the former is slow, they do not need specialised technical facilities, such as the pumps which are required for drilled boreholes.

Gonzales [12] recommended for Mexico the following protective measures:

1. keeping permanent records of groundwater utilisation,
2. definition of restricted zones for controlling the withdrawal and exploitation of groundwater,
3. establishment of a centralised administrative unit responsible for all protective measures,
4. construction of artificial recharge installations whenever economically feasible,
5. protection of groundwater quality,
6. combined utilisation of surface water and groundwater whenever possible,
7. strengthening of the groundwater scientific survey,
8. co-operation between the organisations in charge of the various aspects of groundwater management,
9. reasonable utilisation of groundwater resources.

8.5 ARTIFICIAL RECHARGE

In general, artificial recharge is the practice of increasing the amount of water which enters a groundwater reservoir. The method is based on artificial means and so artificial recharge can be also defined as utilizing surplus water to augment the existing groundwater resources, which may be in a stage of depletion.

Any artificial recharge practised on a large scale and for a prolonged period of time may produce numerous side effects, which in many cases will be negative. Therefore, all the possible future impacts of artificial recharge designed for a specified purpose should be taken into account. In some regions, artificial recharge is used not only for increasing the groundwater resources but also for improving regional management.

Artificial recharge is typically applied to:

1. increase live groundwater storage,
2. replenish the exhausted part of the dead storage,
3. reduce the rate of groundwater level decline,
4. reduce the intrusion of saline or polluted water and/or to improve water quality,
5. dispose of flood water,
6. dispose of and recycle waste water,
7. counteract land subsidence,
8. regulate the groundwater–surface water interaction.

The most common source of water for artificial recharge is a stream, but cooling water, industrial waste water and treated sewage water is also used. The recharging methods can vary depending on, for example, the quality of recharged water and the purpose for which the groundwater storage will be utilised. If some degree of water treatment is necessary, the overall cost of an artificial recharge project may rise considerably.

Recharging methods, will depend also on the purpose of the scheme. According to Jones [17] the immediate hydrogeological environment will determine which of the methods is feasible for the given conditions.

Formerly, the most widely practised method was surface spreading. The method is suitable for unconfined aquifers, and for soils in which the moisture in the zone of aeration can be held above the field capacity for a prolonged period.

In areas where surface infiltration is rather low, recharge pits and lagoons may be preferred. In order to avoid clogging, a basal filter is installed together with facilities which allow the scraping of fine material from the sides and bottom. Obviously more than one pit will have to be installed within a single scheme, to keep the system functioning at those times when one of the pits is being maintained.

For deep and/or confined aquifers, recharge through wells is convenient. In principle, the effect of a recharge well is the opposite of that of an abstraction well. Dry wells, with a perforated cased hole ending above the water table, and wet wells, with the hole below the water table, are recognized.

Sometimes induced recharge is considered as another form of artificial recharge. In such a scheme, the withdrawal of water from wells and infiltration galleries adjacent to the stream will develop an inflow of surface water. The method is limited to unconsolidated formations such as alluvial deposits. The amount infiltrated can be calculated using Jenkins method [14].

In some schemes artificial recharge may become an integral part of large projects. The construction of water gates and dams will cause groundwater level to rise, and the outflow-inflow interaction to change. This type of effect was analysed by Holz [15] in the Weser basin.

Not all types of aquifers are suitable for artificial recharge. Castany [16] considered as suitable karstic plateaux, alluvial plains, littoral dunes, and areas in which the groundwater storage was overexploited. As well as the hydrogeological features, the soil structure and vegetational cover also need to be examined. In areas with thick vegetation a considerable amount of water can be spent on interception and transpiration, and the infiltration rates of some soil types may become considerably reduced after a prolonged period of sprinkling.

8.6 CONJUNCTIVE USE OF WATER

The above term is normally used to the joint use of water from surface and subsurface reservoirs. However, in a broader sense, the joint sequential utilisation of both surface and subsurface water for a certain type of management scheme can also be closely linked with the problems of groundwater and surface water interaction, when any extended use of one water source will sooner or later have an effect on the other one. Only exceptionally is a surface-subsurface water system stable. For instance, in arid zones when live storage forming the baseflow is depleted, there is no precipitation, and the soil moisture in the zone of aeration is close to the wilting point; conversely, perennial streams in temperate regions always interact with the groundwater storage. While groundwater influx into the streams in the form of baseflow is a common feature of many basins, the reverse process is less common and is frequently produced or accelerated by man's activities.

It is not only surface or subsurface water which can be analysed from the point of view of conjunctive use. For instance, Andres and Egger [17] proved in Bavaria that the C^{14} dated Holocene and Pleistocene groundwater in undisturbed areas, and particularly in withdrawal areas, is regenerated by the long-term infiltration of recent groundwater from Neogene and Quaternary aquifer. In the abstraction areas, the infiltration rates are higher, due to an increasing pressure gradient resulting from withdrawal. Thus two different groundwater régimes are involved in a conjunctive use.

The conjunctive use problem is often closely connected with artificial recharge. However, the same problem associated with more or less natural interactions can be of primary importance for economic reasons.

For instance, the temporary recharge of flood water into a shallow aquifer which forms the banks of a river is a good example of conjunctive use. This can happen in areas adjacent to wadis in semi-arid and arid regions. Besbes et al. [18] described the system of groundwater recharge from ephemeral streams in Tunisia. In that region, stored groundwater flows to shebkas – natural depressions with brackish water, and there evaporates. The groundwater level peak occurs 140 days after the flood. The delay is an indicator of the time of travel of the subsurface water, also providing the information that the stored water could be conserved by one means or another. Obviously, in such areas expensive methods of artificial recharge can be avoided by simply using natural recharge.

The induced recharge of groundwater from the stream network is yet another example of a conjunctive process which can be fairly spontaneous or accelerated by human activities. Firstly a considerable amount of water can be obtained from exogenous surface sources, while in the closed system of a single basin the induced water usually comes from the potential baseflow.

Traditionally, the conjunctive use of surface water and groundwater is related to the conjunctive use of surface reservoirs and groundwater aquifers. The surface reservoirs can be natural and/or artificial, while the latter are nearly always of natural origin. Surface water reservoirs usually supply most of the annual water requirements, while the subsurface ones are replenished during years with high precipitation to provide additional water supply during dry years. In very dry years water management may depend on them entirely. For instance, cotton irrigation schemes in southern Zimbabwe depend on the groundwater supply during periods of severe drought, while in normal and wet years water is transported to the fields from distant surface reservoirs in the hills. The tremendous loss which results from transporting water over long distances in dry river beds amounts to 50% of the water released from the reservoirs.

In some regions, the natural recharge of subsurface reservoirs can be accelerated by spreading excess water in ponds and basins specially treated to allow accelerated infiltration and percolation. Recharge pits and recharge wells can also be convenient. The return flow from irrigation schemes can be considered as another important source.

The economic value of conjunctive use can be improved in many ways, for instance, through the probability concept approach as discussed in another section. Listed below are some of the basic possibilities:

1. economic development, in the course of which mainly small reservoirs are constructed and immediately utilised,
2. better timing of water distribution,
3. utilisation of a greater portion of flood water,
4. reduced evaporation from surface water reservoirs,
5. reduced canal lining, based on the presumption that any seepage through the canal banks contributes to the groundwater replenishment.

It should be pointed out that negative effects also can be expected. For instance, Todd [19] lists among the negative effects: less hydroelectric power from small surface reservoirs, decreased pumping efficiency resulting from large fluctuations in groundwater levels, greater mineralisation of water in the subsurface reservoirs, and the danger of land subsidence in confined aquifers.

Conjunctive systems can be successfully applied in arid regions. As an example the system described by Illangasekare and Morel–Seytoux [20] can be mentioned, which is concerned with water management in coastal areas of Saudi Arabia. The surface water is represented by the wadi network, in which surface water resources are ephemeral and most of them would be lost due to a rapid surface runoff of scarce rainfall. The traditional code of water distribution utilises the water from wadis for flood irrigation. Thus the conjunctive system consists of the distribution canals, farm fields, and saturated and unsaturated zones above the aquifer. In the absence of any other water resource, the final objective has to be the improvement of the agricultural techniques in the region.

The conjunctive use of water has reached an advanced stage in many integrated projects. Aspects of effective management based on optimized planning are beyond the scope of this book, and the interested reader will find them in special papers. For instance, the model of Ye Bingru [21] is defined through a number of equality constraints. Kareliotis [22] uses the optimum defined as the minimum cost of the system. Thus the cost of pumping equipment, pumping operations, storage reservoirs and pipes is taken into consideration.

The risk and uncertainty assessment was analysed by Haimes et al. [23]. Their decision-making approach implies:

1. simple or multiple decision-making with some common interest,
2. an agenda or a set of issues represented by a set of multiple objectives which are often non-commensurable, in conflict, or in competition,
3. a set of decisions or policy options restricted by a variety of constraints,
4. an implicit overall goal, shared by the majority of the decision makers involved.

Kisiel and Duckstein [24] simplify the problem of the conjunctive use system so that it provides the possibility of capturing water which would otherwise evaporate. In other words, they attempt to maximize the groundwater recharge by proper pumping in the vicinity of the recharge area. Another aspect (described by the authors as symmetric) is to find a pumping policy which does not cause undue diversion of the surface water that was planned for other users downstream. The authors reject the purely economic approach which tends to prevail in conjunctive use models, because the various alternative approaches to water cycle changes are difficult to measure in financial terms.

A rather hydrological approach to conjunctive use was described by Kemp and Wright [25]. They attempted to regulate river flow by means of the intermittent removal of groundwater. Their aim was to sustain a high river abstraction at a downstream point. They point our that, nevertheless, the riverside amenities, navigation and maximum yield should be sustained. This aim is achieved by redistributing the groundwater storage in time, through an intermittent abstraction of groundwater.

In advanced studies the problems of groundwater management incorporate as well as hydrological phenomena, aspects relevant to the political and sociological environment, legislation and economy. For such cases Schwarz [26] proposed a solution based on the objective function, formulated by:

1. Increasing the yield of the aquifer to the total volume defined as the safe or practical sustained yield (no financial constraints and other sources of water are considered).
2. Minimizing costs for a given supply schedule with specific temporal and spatial distribution of water demands. Here both operational and capital costs should be incorporated.
3. Maximizing net benefits, which are defined as gross benefits minus costs.

In this approach the limitations of the groundwater use are due to physical, economic and legal factors, all of them described as limitations on groundwater levels. As an example of physical limitation we can give salt water intrusion into the adjacent aquifer when the groundwater level declines below a critical point. An economic limitation would be the structure and depth of the wells (again relevant to the groundwater level). Finally, legal aspects consider, for example, the impact of continuous pumping and groundwater level decrease on the groundwater level in the wells of neighbouring owners.

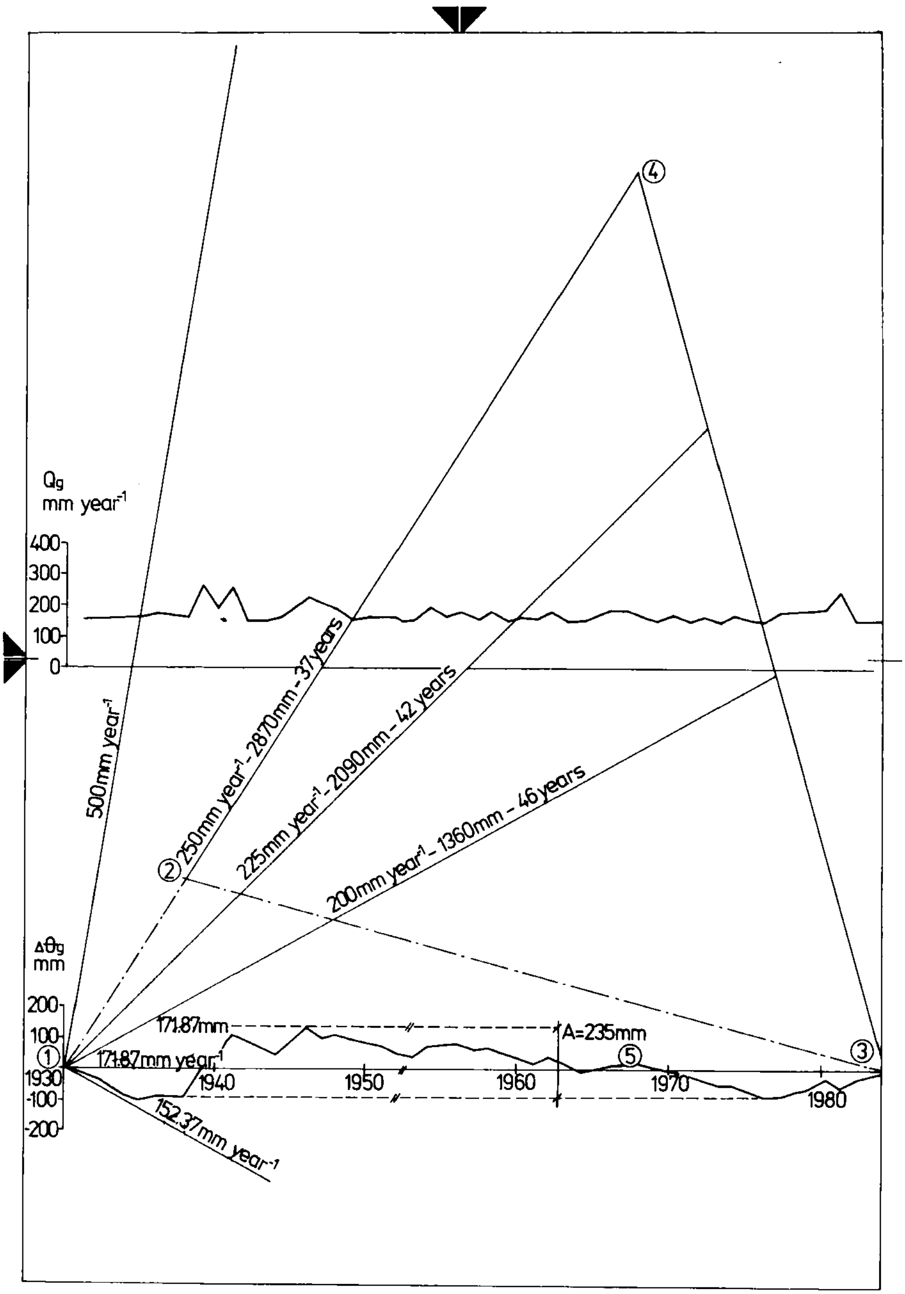
As Biswas [27] stated, interdisciplinary teams should be established in order to solve advanced problems of effective groundwater management, because people from different disciplines can each perceive the same problem in a different light as a result of their training and background.

Stott [28] provides an example of conjunctive water use in the area of Lincolnshire Limestone in England, where the abstraction of the low-cost groundwater could be increased during wetter periods, while during the droughts the pumped storage from the reservoir becomes available. In other regions, of course, a converse solution might be more economical.

O'Mara [29] applies an economic and legal approach to the solution of efficient and sustainable conjunctive water use. He provides several examples to show different approaches should be followed depending on various types of natural and political environment. In general, the assignment of well-defined transferable legal rights, corrective taxes or subsidies, and centralised control of the water resources, is recommended.

Sometimes we may need temporarily to increase the pumping rate, in some cases even in excess of the mean annual baseflow. This is feasible as long as sufficient groundwater is stored in the form of dead storage, and connected somehow with the live storage. A modified mass-curve method can be used in the solution, in the same way as for open reservoir management.

Fig. 8.5 shows a plot of the baseflow (in mm year^{-1}) obtained by conceptual simulation for the period 1931 – 1984. The groundwater régime of the sandstone



8.5 Conjunctive use of dead and live water storage.

aquifer appears to be rather stable. Nevertheless, wet and dry periods can be traced in the mass curve which is drawn so that the mean annual baseflow is identical with axis x and the ordinates are the differences between the mean annual baseflow and the accumulated outflow. The mass-flow ordinates are in mm. The negative ordinates represent those periods when the groundwater outflow is below the mean value of 173.75 mm year^{-1}, and vice versa. The line parallel with the most steeply declining slope of the mass curve represents approximately the safe yield of 152.37 mm year^{-1} which can be pumped throughout the period. At that time of course, very little flow, if any, will occur in the catchment and other environmental damage would follow.

In order to utilise the mean baseflow throught the period we need extra water storage, of which the amount is given by the distance of the "draft lines" parallel with the x axis and drawn tangentially to the highest and lowest point of the mass-curve. An ordinate A equal to 235 mm indicates the amount of water which has to be available in the form of dead storage to supply this water during the dry periods. An equal amount of water will be replenished towards the end of the period.

It should be noted that the draft line does not have to be a straight line. In fact, draft lines in the form of curves are more typical.

More water can be temporarily utilised in aquifers underlined by extensive dead storage reservoirs than it can from open reservoirs. Lines above the axis represent various strategies of water pumping in excess of mean annual baseflow. Let us assume that during the first dry period we may need to pump 250 mm year^{-1} during the subsequent eight years. Thus 2000 mm will be extracted and, because the total outflow during the period is 9381.35 mm, in the following years the mean annual rate of pumping has to be limited to $7381.35/(54-8) = 160.46$ mm. Lines $\overline{1\;2\;3}$ represent that strategy. Because we are limited to the total baseflow volume of 9381.35 mm, the length of the pumping period during which we are allowed to extract 250 mm year^{-1} is $9381.35/250 \doteq 37$ years (point 4). After that period the pumping has to drop to zero. Thus for various strategies we can determine the maximum permissible length of the pumping period and we find that the limits are along the line $\overline{3\;4}$. Of course, for each event another quantity of dead storage must be available. For the maximum pumping of 250 mm year^{-1}, 2870 mm of water has to available in the dead storage, which at a porosity of 10% represents a mean depth of 28.70 metres. That value is found as a distance between the peak point 4 and the mass curve point 5. Naturally, the technical aspects such as the distribution of the wells, the need for a minimum discharge, environmental impacts, possibility of the land subsidence, etc. will have to be taken into account.

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SUBJECT INDEX

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