International Association of Hydrogeologists

David N. Lerner
Arie S. Issar
Ian Simmers

Groundwater Recharge

A Guide to Understanding and Estimating Natural Recharge

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G. Castany, E. Groba, E. Romijn

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PREFACE

At its Sixth Session in Paris, 22-30 March 1984, the Intergovernmental Council of the International Hydrological Programme (IHP) approved the activities to be undertaken during Phase III of the IHP (1984-1989). The overall title of Phase III is 'Hydrology and the scientific bases for the rational management of water resources for economic and social development'. The plan is broadly based, having regard to the varying needs of the developed and developing countries and the fact that the execution of IHP activities in Member States is and will be based on their specific social, economic and cultural patterns.

The plan of Phase III identifies eighteen themes which have been grouped under four main sections. Section I deals with hydrological processes and parameters for water projects, and includes Theme 2.4 'Use of physical and mathematical models for studying the regime of groundwater and predicting changes in quantity and quality'. Project 2.4b produced a short document published in 1987 by Unesco on 'The value of groundwater models for planners and decision makers'. In the framework of Project 2.4c two activities were planned: to hold an international workshop on 'Estimation of natural groundwater recharge' and to prepare and publish a manual of practice on the same subject. Project 2.4d discusses the key issues in groundwater modelling, the representativeness of data and the spatial variability of hydrological variables and parameters. The report 'Consequences of spatial variability in aquifer properties and data limitations for groundwater modelling practice' was published as the International Association of Hydrological Sciences publication no. 175 during 1988.

In 1984 Unesco invited the International Association of Hydrogeologists (IAH) to contribute in two ways to project 2.4c. of its ongoing International Hydrological Programme; viz,

- to hold an international workshop on 'Estimation of Natural Groundwater Recharge'; and
- to prepare and publish a manual of practice on the same subject.

IAH accepted the invitation and in 1985 established a working group under the leadership of Ian Simmers (Amsterdam Free University), with as active members Okay Eroskay (Turkey), Arie Issar (Israel), Gert Knutsson (Sweden), David Lerner (UK) and Erik Romijn (The Netherlands).

From an early stage it was decided by the working group that the target for both activities would be the world's arid and semi-arid zones, these being the areas where the need for reliable estimates of groundwater recharge are greatest. African arid zones in particular have experienced severe droughts during the last decade, causing considerable water supply difficulties, while in a broader (semi-)arid zone framework overexploitation of natural resources, including groundwater, and rapid urban development have created major socio-economic problems.

The international workshop, held in Antalya, Turkey, from 8 to 15 March 1987, was attended by 42 participants from 17 countries and a number of invited observers. Principal organisers were the IAH and Amsterdam Free University (Department of Hydrogeology), with cooperation/support from the University of Istanbul (Engineering Geology section), the State Hydraulic Works of Turkey (DSI), IAEA and IAHS. The workshop was sponsored by NATO Scientific Affairs Division, the US Army European Research Office and Unesco. The Netherlands Ministry of Foreign Affairs provided support funding for several participants from developing countries. Proceedings of the meeting have since been published by Reidel in the NATO ASI-Series C as vol. 222 (1988).

The present 'manual of practice' reflects working group opinions and concerns with respect to the state-of-the-art of arid/semi-arid area groundwater recharge estimation. The volume is not intended as a 'cookbook', but offers guidance to the practitioner engaged in arid and semi-arid zone water resources exploration and development. The volume is in four parts:

Part I introduces the study framework, defines the recharge concepts which follow, and discusses the problems of space and time variability in relation to the translation of point measurements to regional recharge estimates. Brief comments are also given on some of the implications for resource management.

Part II deals with the concept of hydrogeological provinces. Characteristics and case studies are summarized and discussed for each, thus developing a series of typical hydrogeological conceptual models, the section ending with a general procedure algorithm for deriving 'first estimate' recharge values.

Part III gives details and examples of techniques currently available for quantifying groundwater recharge. Specifically considered are direct measurements, water balance and soil moisture budgetting methods, Darcian approaches, tracer techniques and empirical procedures. Each is comprehensively and critically evaluated with respect to recharge source (precipitation, rivers, interaquifer flows, irrigation and urbanisation), with summaries for ready reference.

Part IV closes the volume with a series of case studies chosen to illustrate the use of various techniques in a number of hydrogeological provinces. Since the examples given are largely abridged versions of papers already contained in the Antalya workshop proceedings, the authors and working group acknowledge the willing cooperation of the

publishers (D. Reidel, Dordrecht) in making the material available.

Numerous individuals and organisations have been involved either directly or indirectly in the preparation of this manual. Financial support was provided by Unesco, the Swedish IHP National Committee and in fellowship form to one of the authors by the Jacob Blaustein Institute for Desert Research (Ben-Gurion University of the Negev, Israel). Also gratefully acknowledged are the personal contributions from Ron Passchier (Part II), Dr. S.D. Limaye (Part II, volcanic terrains), Robert Gray (Part III) and Ersin Seyhan. Per-Olof Johansson (Sweden) served as technical editor for the volume and valuable review advice was received from Dr. D.C.H. Senarath (University of Moratuwa, Sri Lanka), Dr. E.S. Simpson (University of Arizona, USA) and the Hydrological Branch of IAEA. Mr. Nelson da Franca, Programme Specialist of the Division of Water Sciences of Unesco, was the responsible person within the IHP Secretariat for this project. The manual would not have been possible without these various forms of support and the considerable team effort by an active working group.

My personal appreciation, as Chairman, goes also to the IAH Council, to the principal authors, and in particular to Erik Romijn for his dedication to scientific and administrative detail.

Ian Simmers
Chairman,
Working Group on Estimation of Natural Groundwater Recharge

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GROUNDWATER RECHARGE

A guide to understanding and estimating natural recharge

Part I : ARIDITY, GROUNDWATER RECHARGE AND WATER RESOURCES MANAGEMENT

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1 ARIDITY, GROUNDWATER RECHARGE AND WATER RESOURCES MANAGEMENT

In this volume on natural and 'man-induced' groundwater recharge estimation the principal concern is with how best to quantify and hence model the process, with a view to providing a more rational approach to overall water resource management. Target areas for the discussions and outlines of the numerous available procedures which follow are the world's arid and semi-arid zones.

It must be stressed at the outset that the information contained represents an appraisal of the present state-of-the-art with respect to arid/semi-arid area recharge determination and does not aspire to being the ultimate word on the subject. Standard procedures for the collection and processing under arid or semi-arid conditions of such hydrological variables as precipitation, evapotranspiration and streamflow are not described in detail; for these reference should be made to one of the various text- or handbooks (e.g. FAO, 1981). Finally, the volume is not intended as a 'cookbook' on the general problems of recharge estimation. It does not, therefore relieve the reader of the need for independent thought on a specific problem, but should be considered as a source of information to facilitate a logical and structured approach to the steps involved.

Initial motivation for this manual of practice is that one third of the world's land surface has been classified as arid or semi-arid and approximately half the countries are directly affected in some way by problems of aridity (UNESCO, 1977, 1979). Easily developed land has in large measure already been exploited and attention is thus increasingly towards more arid areas for human survival. However, soil and water resources of arid and semi-arid regions are limited, often being in a delicate environmental balance. Surface water supplies are normally critically unreliable, poorly distributed and subject to high evaporation losses. For the rapidly expanding urban, industrial and agricultural water requirements in these areas, groundwater use is thus of fundamental importance. This in turn creates a host of associated problems since, for example, abundant available groundwater may have only small natural recharge, thus raising such issues as mining a non-renewable resource, quality deterioration by saline water intrusion and land subsidence.

Quantification of the current rate of natural groundwater recharge is thus a basic prerequisite for efficient groundwater resource management, and is particularly vital in arid regions where such resources are often the key to economic development (Foster, 1988). Unfortunately, of all the factors in the evaluation of groundwater resources, this rate of aquifer replenishment is one of the most difficult to derive. Equally true is that it usually takes time for social awareness of a new problem to be aroused and effective controls initiated. By this time groundwater may be exhausted and large capital investments lost. It may therefore be concluded (FAO, 1981) that 'a higher level of competence, not only technical but social and political, is needed in arid than in humid zones to achieve sustained progress'.

These above factors serve to illustrate the growing international demand for reliable quantitative information on arid and semi-arid zone groundwater recharge estimation and hence form the catalyst for the present manual.

1.1 Definitions

1.1.1 Aridity

Although terms such as aridity are somewhat vague, with any classification influenced by the intended use, a number of environmental features characterise the so-called arid and semi-arid areas of the earth (FAO, 1981):

- 'high levels of incident solar radiation;
- generally high diurnal and seasonal temperature variations;
- low humidity at short distance from the sea;
- strong winds with frequent dust and sand storms;
- sporadic rainfall of high temporal and spatial variability;
- extreme variability of short-duration runoff events in ephemeral drainage systems;
- generally high infiltration rates in channel alluvium;
- high sediment transport rates;
- relatively large groundwater and soil moisture storage changes;
- distinctive geomorphology, with negligible weathering processes and poorly developed soil profiles'.

The essential characteristic, however, and the one upon which all others depend, is the smallness of precipitation.

Principal factors causing aridity, either singly or in combination, are summarised by UNESCO (1977) and Lloyd (1986) as:

- the high pressure belts located at sub-tropical latitudes, giving hot, dry subsiding air;
- continentality and coastal mountain rain-shadow effects;
- the effects of cold oceanic currents along some coasts.

All these factors are evident in the readily available maps of arid and semi-arid areas presented by UNESCO (1953, 1958, 1979) and as shown in simplified form by Fig. 1.1 taken from Hodge and Duisberg (1963).

Precise definitions of aridity remain difficult, despite the many attempts found in the literature, because of the variety of climatic and other data (e.g., topography, soil condition, history and type of landuse) which must be taken into account.

A selection of these based solely on annual average precipitation (mm) is as follows:

Hyperarid	Arid	Semi-arid	Reference
0-50	50-200	200-500	Lloyd (1986)
-	< 100	100-500	G. Droughin (UNESCO, 1953)
0-200	200-400	400-800	Y.M. Simaika (UNESCO, 1953)

The diversity of these values indicates that total precipitation is not a sufficient definition and that such aspects as rainfall duration, length and timing of the dry season and some measure of temperature or water availability for plant growth also need to be considered for the present manual.

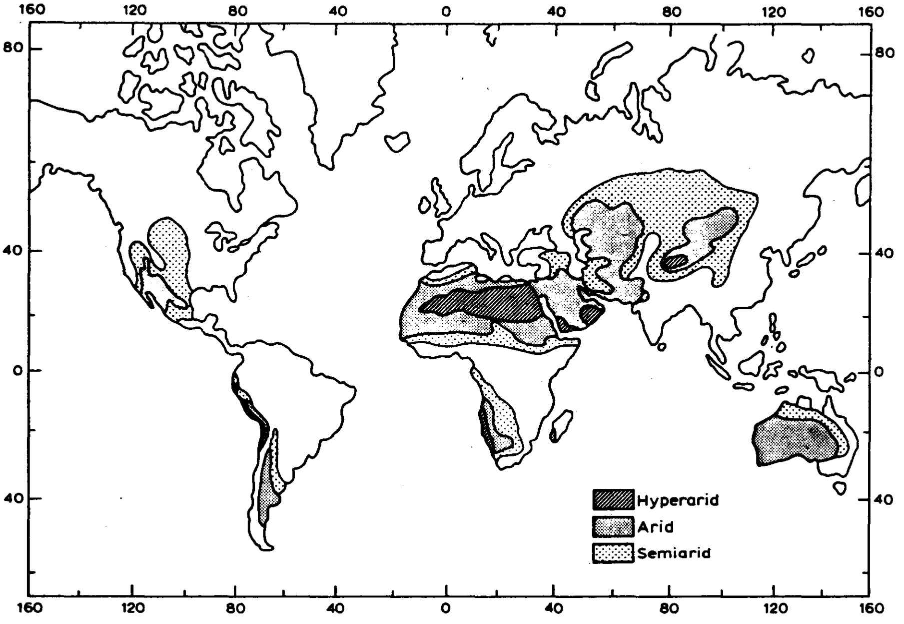

Fig. 1.1 Arid and semi-arid areas of the world (Hodge and Duisberg, 1963)

Failing a more comprehensive classification therefore, that presented by UNESCO (1979), based on the earlier work of Meigs (1953), is preferred for general use. In this, the delimitation of hyperarid, arid and semi-arid areas is based on aridity indices and all available data on soils, relief and vegetation. Full details are given in the UNESCO maps and accompanying explanatory notes, a brief definition of each being:

Hyperarid zone ($p/et_p < 0.03$, where p and et_p are respectively mean annual precipitation and potential evapotranspiration), annual rainfall is very low with interannual variability up to 100%, very sparse vegetation and no rainfed agriculture or grazing.

Arid zone ($0.03 < p/et_p < 0.20$), annual rainfall is 80-150 mm and 200-350 mm in respectively winter and summer rainfall areas, interannual rainfall variability 50-100%, scattered vegetation, nomadic livestock rearing is possible and agriculture based upon local rainfall is only possible through rainwater harvesting techniques.

Semi-arid zone ($0.20 < p/et_p < 0.50$), annual rainfall is 200-500 mm and 300-800 mm in winter and summer rainfall areas, interannual variability 25-50%, discontinuous vegetation with perennial grasses, rainfed agriculture and sedentary livestock rearing are common.

1.1.2 Groundwater recharge

Groundwater recharge may be defined in a general sense as the downward flow of water reaching the water table, forming an addition to the groundwater reservoir. A clear distinction should thus be made, both conceptually and for any modelling purposes, between the potential amount of water available for recharge from the soil zone and the actual recharge as defined above. Rushton (1988) shows that the two quantities may differ, due to either the influence of the unsaturated zone or non-acceptance by the aquifer of the potential value. This aspect is discussed further in Section 11.1.2.

Recharge of groundwater may occur naturally from precipitation, rivers, canals and lakes and as a man-induced phenomenon via such activities as irrigation and urbanisation - losses from irrigation programmes frequently provide a contribution which exceeds that from rainfall.

Two principal types of recharge are recognised, categorised here (Fig. 1.2) as direct and indirect (FAO, 1981; Lloyd, 1986). Other terms which have been used as equivalents are local (or diffuse) and localised recharge (see, for example, Allison, 1988; Foster, 1988).

Direct recharge is defined as water added to the groundwater reservoir in excess of soil moisture deficits and evapotranspiration, by direct vertical percolation of precipitation through the unsaturated zone.

Indirect recharge results from percolation to the water table following runoff and localisation in joints, as ponding in low-lying areas and lakes, or through the beds of surface watercourses. Two distinct categories of indirect recharge are thus evident; viz, that associated with surface water courses, and a second localised form resulting from horizontal surface concentration of water in the absence of well-defined channels (see also sections 11.1.1, 11.8 and Chapter 12).

These definitions are of course a simplification of reality, since lateral subsurface recharge is not explicitly considered, so-called preferred pathways are a common phenomenon with even direct recharge (Sharma and Hughes, 1985;

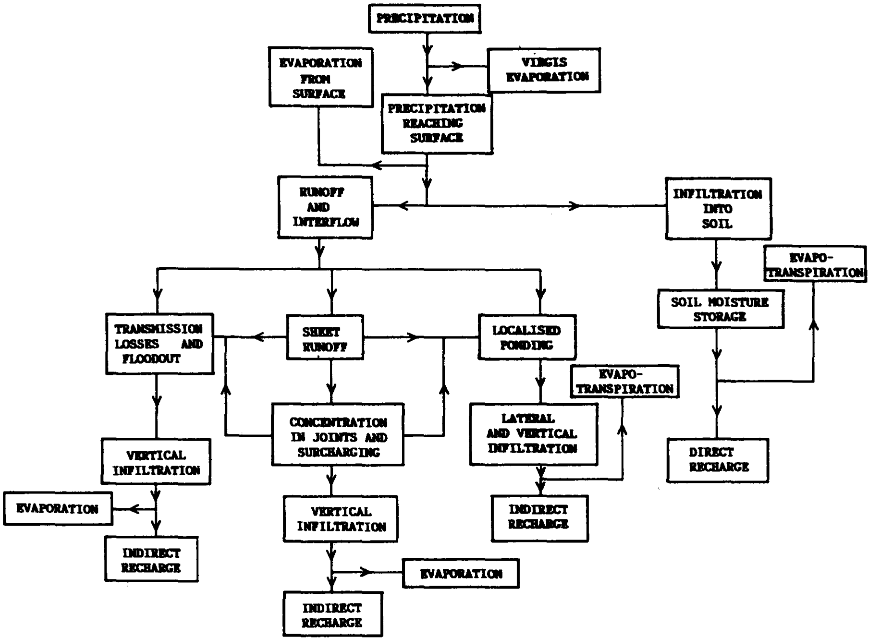

Fig. 1.2 The various elements of recharge in an arid area (Lloyd, 1986)

Johnston, 1987) and in many locations a combination of both direct and indirect recharge, as defined, will occur. However, for modelling purposes a number of general guidelines are evident from the international literature:

- there is no doubt that recharge occurs to some extent in even the most arid regions, though increasing aridity will be characterised by a decreasing net downward flux and greater time variability;

- as aridity increases, direct recharge is likely to become less important and indirect recharge more important in terms of total recharge to an aquifer;

- estimates of direct recharge are likely to be more reliable than those of indirect recharge.

These generalisations indicate that successful groundwater recharge estimation depends on first identifying the probable flow mechanisms and important features influencing recharge for a given locality, since it cannot be assumed that a procedure successfully developed for one area will prove equally reliable for another.

Rushton (1988) lists several of the factors affecting recharge as follows:

At the land surface:

topography
precipitation: magnitude, intensity, duration, spatial distribution
runoff, ponding of water
cropping pattern, actual evapotranspiration

Irrigation:

nature of irrigation scheduling
losses from canals and water courses
application to fields, land preparation, losses from fields

Rivers:

rivers flowing into the study area
rivers leaving the study area
rivers gaining water from or losing water to the aquifer

Soil zone:

nature of the soil, depth, hydraulic properties
variability of the soil, spatially and with depth
rooting depth in soil
cracking of soil on drying out or swelling due to wetting

Unsaturated zone between soil and aquifer:

flow mechanism through unsaturated zone
zones with different hydraulic conductivities

Aquifer:

ability of the aquifer to accept water
variation of aquifer condition with time.

The actual frequency of recharge events and the transit time until recharge takes place are also important, differences obviously influencing both the choice of method for recharge estimation and eventual resource management. For example, infrequent major recharge is a totally different proposition from smaller but more regular events (Fig. 1.3), since in the former case the negative side-effects of overdevelopment may have already occurred prior to the replenishment (Foster, 1988). Further, the existence of extensive well-established aquifer hydraulic gradients is no guarantee of recent recharge. Such features may largely represent paleo-recharge (fossil gradients), with natural recession of groundwater continuing to the present day (De Vries, 1983; Lloyd, 1986).

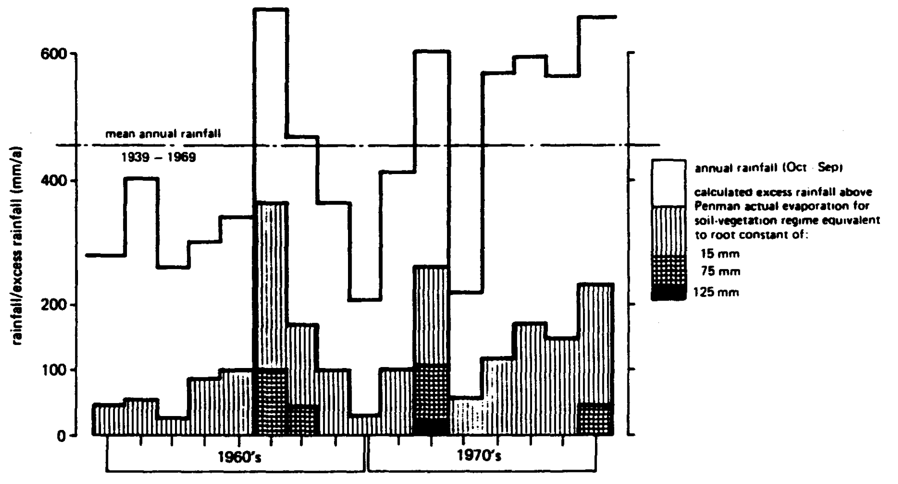

Fig. 1.3 Summary of analysis of daily meteorological data for a site in the Botswana Kalahari (Kweneng District) during 1961-77 (Foster, 1988)

Note: Infiltration rates and frequencies exhibit marked sensitivity to variation in soil moisture-vegetation regime which is in practice laterally variable and difficult to quantify, but groundwater table fluctuations suggest that widespread recharge was limited to a single rainfall episode in the 1971-72 wet season.

1.2 Groundwater recharge processes

Lloyd (1986) suggests that 'hydrological processes in arid areas are no different from those under other climatic regimes, except that in some circumstances the interrelationships between processes are more accentuated under arid conditions and the amounts involved in a process frequently more extreme'. Given a vegetated area this is certainly true for the processes of groundwater recharge, as shown in a relatively simplistic manner by Fig. 1.2. However, a major difficulty in arid areas, thus prompting the present manual, is that although basic recharge mechanisms are reasonably well known, deficiencies are evident in quantifying the various elements. Considering the general scarcity and variability of hydrogeological data in most arid and semi-arid zones, this is to be expected.

It is clear from Section 1.1.2 that differences in sources and processes of groundwater recharge will mean that the applicability of available estimation techniques will also vary. To proceed from a well-defined conceptualisation of the various recharge processes is thus essential.

Although direct recharge is known to be of decreasing significance with increasing aridity, the processes involved are conceptually the easiest to define and form the basis of numerous recharge estimation techniques currently in common use (see Chapter 11). Assuming a dominant vertical moisture flux, a single porous medium and a water table which is not close to the surface, water is postulated to move by Darcian flow in the unsaturated zone to the groundwater body. Since both upward and downward fluxes can occur, for recharge estimation purposes interest lies in the actual steady-state net moisture transfer to the water table. Flow equations representing this net process are not given here and reference should be made to either Section 11.5, Rushton (1988), Johansson (1988), or one of the numerous standard texts on soil physics (e.g., Hillel, 1982).

However, field experiments show that volumetric water content and flow mechanisms in the unsaturated zone vary in a complex manner, the main problem being that the parameters moisture content, matric potential and hydraulic conductivity are sensitively interrelated. For example, a change in the volumetric water content of 5% often corresponds to a change in the hydraulic conductivity by two or more orders of magnitude (Rushton, 1988). Further, material in the unsaturated zone rarely displays homogeneous properties, often consisting of layered sands, silts and clays with widely varying saturated hydraulic conductivities (Fig. 12.2), and a strong potential for lateral rather than vertical flow above lithological discontinuities. The irregular occurrence of preferred pathways in even relatively homogeneous material is an added complication for recharge estimation, as are the problems arising from shallow water tables and from process space and time variability common to arid and semi-arid areas. This last aspect is dealt with further in Section 1.3.

The theoretically simple processes of direct groundwater recharge are thus by no means easy to quantify in nature, and unfortunately the task is even more difficult for either category of indirect recharge. Using Fig. 1.2 as an indicator of indirect recharge mechanisms, Lloyd (1986) concludes that:

- 'the balances between runoff to joints, hydraulic surcharging of materials in joints and evaporative losses coupled with slope, ground surface conditions and erratic precipitation, pose (considerable) problems for hydrological analysis;

- indirect recharge through depression ponding is equally difficult to evaluate as accumulations of clay restrict vertical infiltration while lateral infiltration into adjacent materials is only active with initial flooding;

- transmission losses during flooding are also difficult to quantify. Flood volume differences between two gauging stations are often of the order of measurement error, while any infiltration is subject to high evaporation losses that are impossible to determine'.

In summary, quantification of groundwater recharge is fraught with problems of varying magnitude and hence substantial uncertainties. To reduce result uncertainty it is therefore desirable to apply and compare a number of independent approaches, as allowed by available data.

1.3 Recharge time and space variability

Variations in groundwater recharge with time and in space (both laterally and vertically) are well documented and are the direct consequence of such factors as differing precipitation, soil characteristics, vegetation, landuse and topography.

Given this variability, the obviously interrelated question is what techniques should best be used to derive reliable recharge estimates? Although aspects of this issue are addressed in detail by subsequent chapters, it is clear from the literature that not all methods are strictly comparable in terms of their applicable space and time scales. Some are intended to estimate recharge over an area for long time periods, while others are concerned with short time interval point information.

Table 1.1 Factors influencing the choice of time step for recharge estimation (Lerner, 1987, pers.comm.)

•Time period over which recharge is averaged:	Instantaneous	Event	Season	Year	Historical average	Geological time
•Time steps used for:						
-Scientific studies:	|———	———	———	———	———	———|
-Groundwater resource studies:		|———	———	———	———|	
•Time steps for various factors in a resource study:						
-size of study area:		small ←	—	—	→ large	
-level of study:		design ←	—	—	→ reconnaissance	
-degree of aridity:		arid ←	—	—	→ humid	
-resource exploitation:		heavy ←	—	—	→ light	
-quantity of data:		large ←	—	—	→ small	

Factors which influence the choice of time scale are shown in Table 1.1 - the study objective is clearly of prime importance. From an operational point of view, Table 1.1 should be viewed in conjunction with Fig. 10.1, which shows that although a particular recharge estimation method may be applicable over a spectrum of time scales, some intervals are more directly appropriate than others. This aspect can be quite critical when interest lies in areas with erratic or clustered rainfall, potentially low values of recharge and limited aquifer storage.

1.3.1 Extrapolation in time

Arid and semi-arid zone precipitation is characterised by high interannual variability (Section 1.1.1), as illustrated by Fig. 1.4, resulting in very variable processes over a long time scale. This in turn can lead to considerable recharge estimation problems if long-term values are required with only short-period data available, particularly in the situation of high and/or increasing groundwater exploitation.

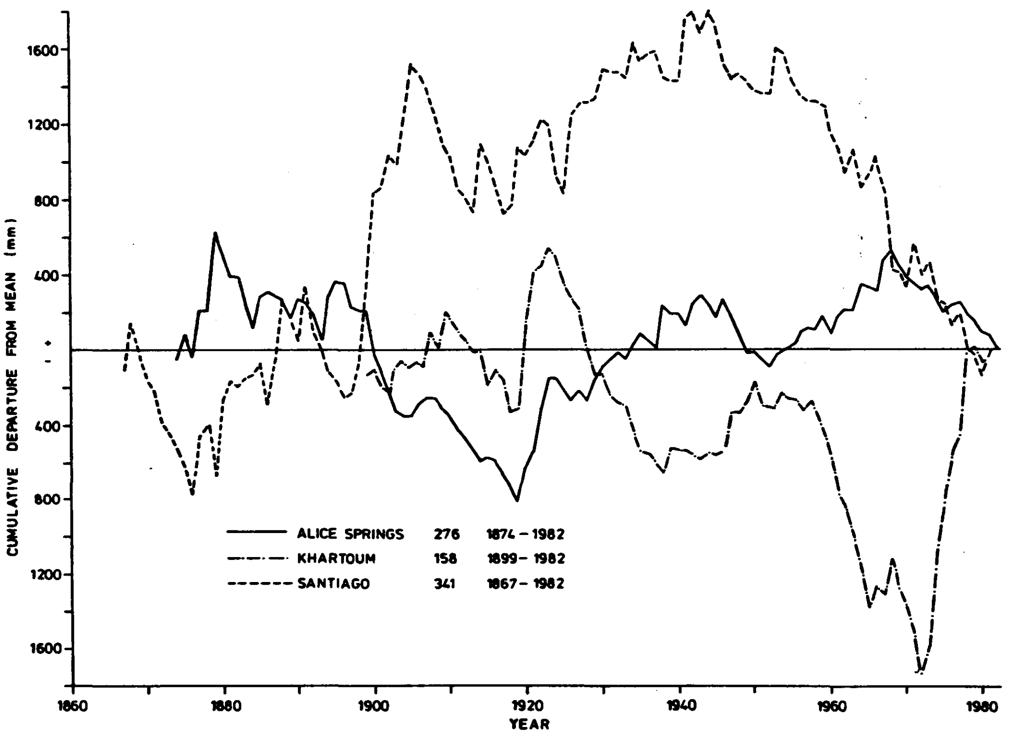

Fig. 1.4 Precipitation trends as cumulative deviations from the mean for example stations in or adjacent to major arid areas. Period mean annual values (mm) shown (Lloyd, 1986)

A possible solution to the problem is to model recharge using stochastically generated long-term rainfall sequences as basic input. For appropriate techniques the reader is referred to Kelman (1977), Beven (1986, 1987) and any of the readily available texts on stochastic modelling.

Although recharge estimation procedures generally assume precipitation and evapotranspiration to be the only time-variant factors affecting the various processes, most techniques can accommodate changes in landuse over the duration of data collection provided recharge response time lies within the period of records. However, if interest is in predicting the consequences for recharge of landuse modifications over time, and hence changes in both vegetation and soil hydraulic characteristics, then a new dimension of complexity is introduced. The obviously related problem is the situation where transit time until recharge is long and landuse changes are known to have occurred in an area prior to current data collection programmes (see also Section 1.4).

Reliable answers to these common management problems are not easily obtained, since the hydrological system is in a state of dynamic evolution, with individual elements of the system characteristically displaying different time frames of adjustment. Current research (e.g., Peck and Williamson, 1987) indicates that the most realistic approach to a solution involves a combination of intensive data collection, translation of nearby or other relevant supporting information and modelling. Unfortunately, rigorous adherence to such an approach, in an attempt to minimise result uncertainty, can be both time consuming and expensive. Some relaxation of pre-modelling data collection is of course always possible, but will in general lead to greater (and perhaps unacceptable) estimation errors.

1.3.2 Spatial extrapolation

As concluded by Allison (1988), probably the most important problem to be overcome in the estimation of groundwater recharge is the assessment and prediction of its spatial variability. Over some quite large areas it appears to show little lateral variability, while in other, apparently similar areas it can range over at least an order of magnitude - numerous examples of this phenomenon are described in Peck and Williamson (1987).

Fig. 1.5 is typical, and shows four profiles of chloride concentration in soil water from what appears to be a uniform field with sand/sandy loam surface soils. Allison (1988) believes that the large variations in chloride concentration reflect correspondingly large changes in recharge rate, the difficulty being that in other apparently similar sites chloride profiles are uniform.

Whatever the reason for the displayed variability, its field identification, quantification and development of statistical or other techniques for estimating recharge over an extended area remain a problem.

This issue has considerable practical importance. If, for example, recharge estimation is based on point measurements, then water resource managers will need to know how or whether these relate to values over a specified area of interest. Unfortunately the cost of multiple point recharge measurements is usually prohibitive, particularly if interannual recharge variability is also high and values are required over an extended time period. A descriptive insight into recharge spatial variability is of course available by way of standard hydrogeological mapping techniques (e.g., borehole logs and pumping tests, hydrochemical sampling, geophysical survey, aerial photograph and thematic map interpretation). However, to create such a map is time consuming and the end product can at best only indicate the general pattern of likely recharge areas and groundwater flow lines. The quantitative information ultimately required for modelling and resource management is not produced by this approach.

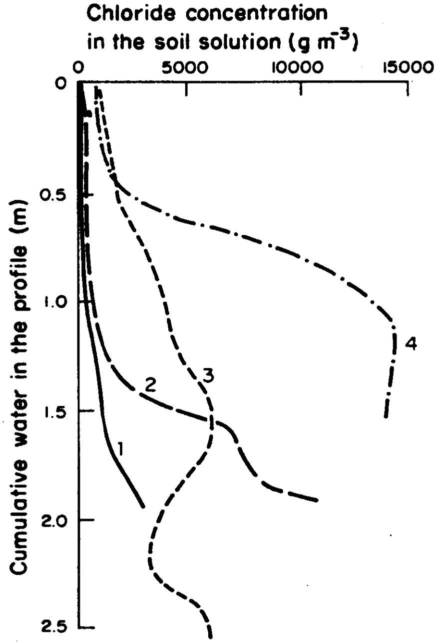

Fig. 1.5 The relationship between chloride concentration of soil water and cumulative water stored in the profile in a field, cleared about 50 years ago, under a pasture-crop rotation (Allison, 1988). All sites are within a radius of 100 m. Rainfall is ≈ 300 mm/a

Strategies to cope with reliable quantitative estimation and prediction of spatially variable recharge have yet to reach a stage of practical application, though current research indicates some potential via a combination of selective ground truth data collection, geostatistics (autocorrelation and kriging) and remote sensing. Details of ongoing activities lie outside the scope of this manual introduction, but a series of interdependent sequential steps can be identified:

- determination of significant physiographic and hydrological parameters which may be used as simplifying surrogates to define the dynamics of the hydrological system (see Table 11.4);

- hydrological response unit identification by way of, for example, geometric grid cells and statistical analyses (Sivapalan and Wood, 1986; Hendriks, et al., 1987);

- 'regionalisation' of the digitised grid cell data using hydrologically oriented classification procedures, with subsequent application of specific recharge techniques and eventual models in each of the classified units.

A systematic approach of this type, based on the spatial scale of variability for measurable physical characteristics, will minimise areal recharge estimate uncertainty for a given expenditure on data collection. However, until these techniques have sufficiently evolved to be of value for practical model application, it would appear that to resolve the spatial variability problem there is little alternative to multiple site recharge estimation in the area of interest. These thoughts are developed further in the suggested procedure algorithm presented in Chapter 9 and in Section 11.7.

1.4 Resource management implications

An introduction to a manual on arid and semi-arid zone groundwater recharge estimation would be incomplete without some comment on associated implications for water resources management. In reality the issue is complex and for optimum value should be site specific, though a number of generalisations are evident from the literature. This section should thus be considered as a summary overview and draws heavily on the more detailed information presented by Lloyd (1986) and Foster (1988).

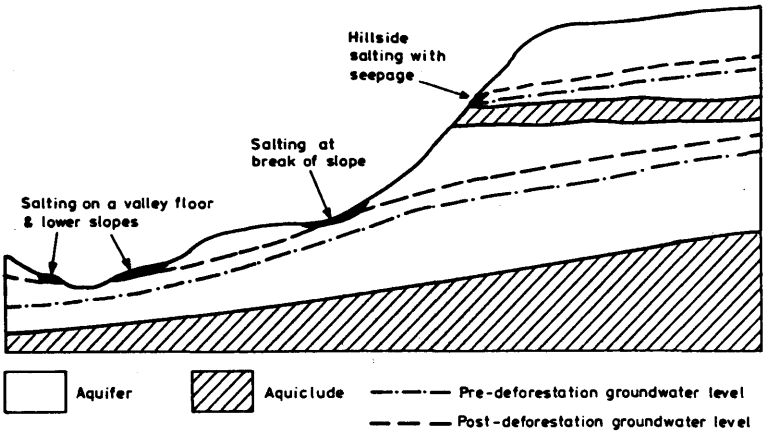

Fig. 1.6 Concepts of salination down the hydraulic gradient into progressively arid areas of Victoria, Australia, as a result of deforestation in the recharge area (Lloyd, 1986)

Two major recharge connected problems in arid and semi-arid area groundwater development are groundwater quality deterioration and resource overexploitation. An obvious example of the first problem would be the too frequently recorded increases in dissolved solids associated with poor agricultural and irrigation practices. Less immediately obvious are the long-term consequences of landuse changes in groundwater recharge areas. Fig. 1.6 (from Lloyd, 1986) is an Australian example of this situation, where severe dryland salination over large areas is a direct result of increased recharge following deforestation in the 19th century.

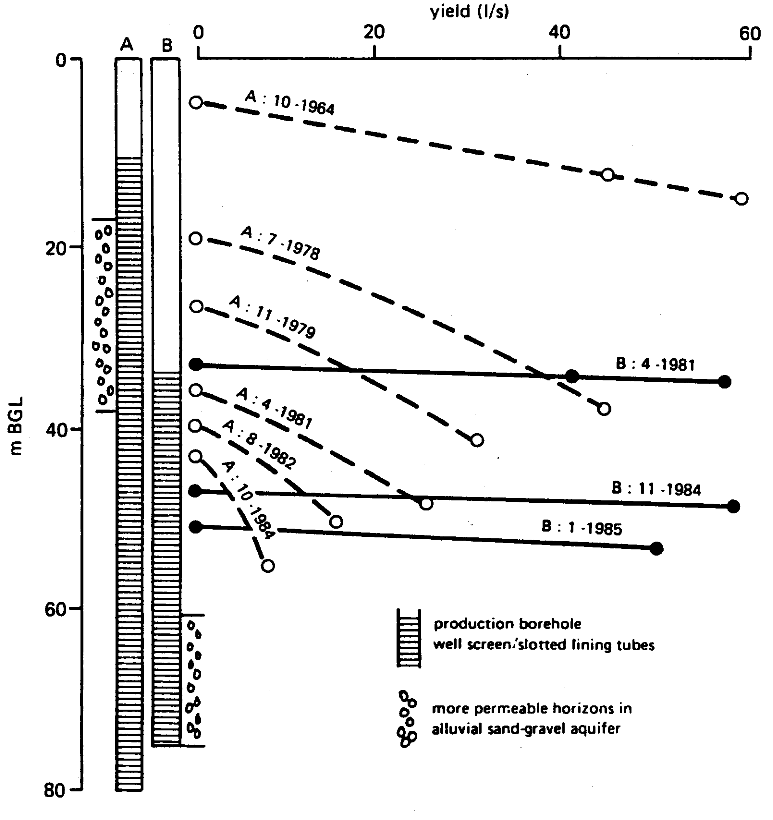

Fig. 1.7 Production performance of pumping boreholes in the overdeveloped alluvial fan aquifer of Lima, Peru (Foster 1988). The contrasting behaviour reflects the falling groundwater level and varying depth at which permeable horizons occur.

Since arid and semi-arid region recharge estimates are often subject to considerable uncertainty and large error, groundwater overdevelopment can occur if active recharge is overestimated. For example, it is not uncommon to find that model based resource evaluations initially overestimate recharge, the classical pattern shown by increasingly refined studies of an area being that of lower recharge values and hence smaller groundwater resources. A number of the consequences of recharge overestimation are listed by Foster (1988) to be:

- 'increased pumping costs, yield reductions and even complete failure of production boreholes (Fig. 1.7);

- the encroachment of saline water into freshwater aquifers in some coastal and inland basin situations;

- land subsidence consequent upon settlement of under-consolidated lacustrine, deltaic or estuarine sedimentary aquifers'.

However, groundwater mining can be an acceptable practice in arid areas provided it is carefully planned and controlled and subject to realistic ongoing evaluation.

A further groundwater resource development problem, particularly in developing countries, is that there are frequently inadequate data to permit reasonable calibration of any recharge model, thus introducing constraints to the assessment of management options. Since large-scale collection of additional data prior to project implementation is usually impractical for reasons of cost and required development timetable, both Lloyd (1986) and Foster (1988) strongly advocate adopting a flexible approach to project design and management. This implies that groundwater resource development should be staged, allowing progressive aquifer response data collection and resource evaluation. A further practical reason for such an approach is that understanding of the hydrogeological system is sometimes difficult until it is stressed. A conceptual framework for such a programme is illustrated by Fig. 1.8 and an example is given in Fig. 1.9.

Foster (1988) does state, however, that under some circumstances this approach will have a more limited application; viz, where

- 'the minimum viable first stage water demand is very large relative to exploitable aquifer storage;

- the profitability of proposed groundwater use is highly sensitive to energy costs;

- there is a significant risk of saline encroachment in an aquifer as a consequence of medium-term overdevelopment'.

1.5 Recharge requirements for groundwater modelling

It is clear from previous sections that groundwater recharge can be modelled at many scales for many different purposes. The required information on recharge varies in each case, that for pollution transport studies within a village on a hard rock aquifer being different from a groundwater resources project for a regional aquifer of 1000 km^2.

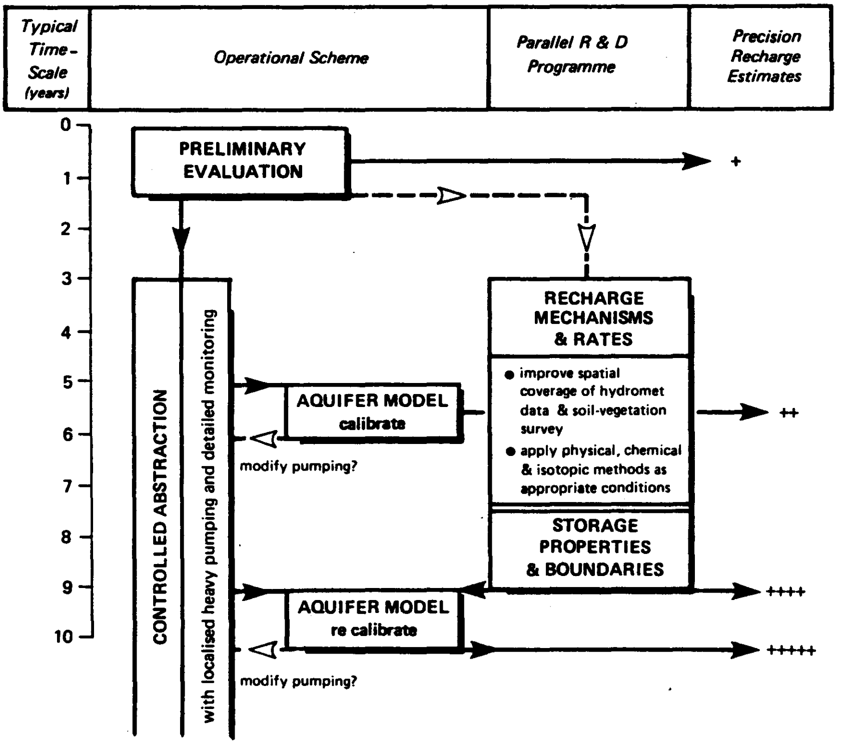

Fig. 1.8 Outline scheme for the integrated operational approach to improving estimates of aquifer recharge rates and storage parameters (Foster, 1988)

For resources studies interest is generally in both the total amount of recharge (since total recharge = replenishable resources) and its spatial and temporal variability (because the optimum use of resources depends upon their distribution). The amount of detail required about recharge variations is in turn governed by the degree to which groundwater resources are (to be) used, and upon the nature of the recharge processes. This means that if only 5% of the resources are used, less detail is needed concerning temporal variations; the aquifer storage will carry through a drought, and a modelling study using long-term average data will suffice.

There is also little point in regional models estimating recharge variation across areas of less than several nodes. In reality however, and for the sake of estimate precision, non-linearities of the recharge process may require values to be determined in more detail than would otherwise be needed for a model, with subsequent averaging over space and time prior to any simulation. Unfortunately data are often both scarce and short-term in arid and semi-arid areas, resulting in an erroneous spatial picture and recharge estimates which may be completely unrepresentative of the long-term pattern (see Section 1.3). In such instances the direct use of groundwater information can improve recharge determinations.

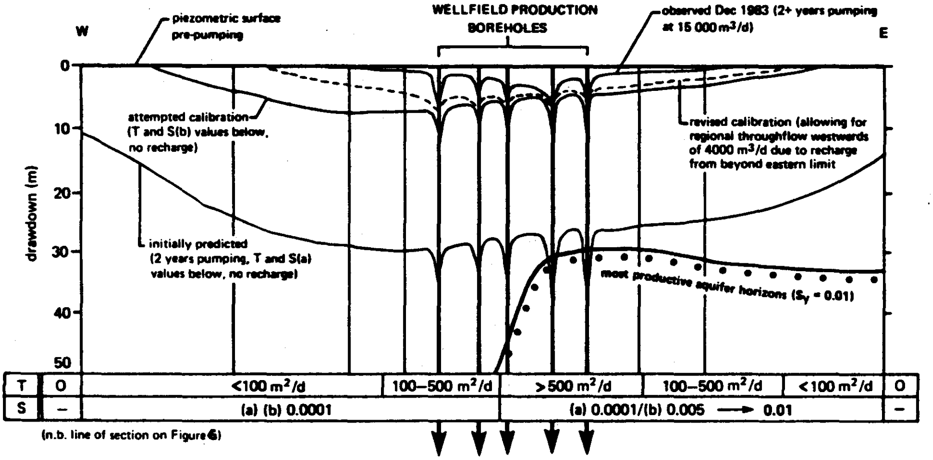

Fig. 1.9 Schematic cross-section of the Jwaneng northern wellfield, Botswana Kalahari, showing initital prediction of the aquifer numerical model and the calibration based on operational monitoring experience with revised storage and recharge parameters (Foster, 1988)

There are of course some cases when there will be minimum value in estimating groundwater recharge; for example, in a limited life scheme where demand will greatly exceed replenishable resources. In this case aquifer storage is of much greater importance than recharge and the system can be conservatively modelled with a zero recharge component. All efforts should thus go to estimating specific yield; any recharge which does occur will provide a safety factor towards the end of the project's life. A second case would be when recharge is low compared to throughflow. The uncertainties in geology, aquifer boundaries and properties, groundwater levels and even borehole abstractions are so great that although a few millimetres per year of recharge may be of great scientific interest, they will be of little water resources concern. On the other hand, in some instances even very small increases in groundwater recharge can have an important practical impact on other aspects of arid zone hydrogeology (e.g., (re-)activation of local springs).

Although the dangers of generalisation are obvious, an element in common to all water resources development decisions involving recharge is the need to initiate investigation programmes on the basis of a <u>hydrogeological conceptual model</u> - defined here as a description of the hydrogeological conditions including the distribution and properties of rocks and the flow of water. Such a conceptual model is the key to logical subsequent development of the most appropriate mathematical/numerical techniques.

1.6 Conclusions

Reliable estimation of groundwater recharge in arid and semi-arid areas is neither straightforward nor easy. Although quantitative information on recharge is often critical for optimum resource modelling and management, no single comprehensive estimation technique can yet be identified from the spectrum of methods available. Differences in sources and processes of recharge mean that the applicability of available procedures will differ, with significant practical implications attached to whether the derived results apply at a point or over a wider area.

The need to proceed from a well defined conceptualisation of different recharge processes is thus necessary, as is the need to use more than one technique for result verification - recharge estimation should thus be viewed as an iterative process, not as a once-and-for-all calculation. As concluded by Knutsson (1988), final choice of techniques will also be determined by the study objectives, initial data base and the possibilities to supplement this, and of course available financial resources.

Despite the spectrum of problems associated with groundwater recharge estimation, the chapters which follow guide both the general reader and practitioner in a structured fashion through the maze of pitfalls and options involved.

Part II : REGIONAL HYDROGEOLOGICAL CONCEPTS

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2 HYDROGEOLOGICAL PROVINCES - INTRODUCTION

Water, on its way from the surface to the aquifer, has to pass a series of obstacles which tend either to return it to the atmosphere or to retain it in the unsaturated zone. One can view the processes on the surface and in the unsaturated zone as a system of pumps which act on the rainwater after it touches the land surface. Each pump is activated by another type of energy; for instance, evaporation can be regarded as a thermodynamic pump, transpiration mainly as an osmotic pump connected to a thermodynamic one. Another set of pumps causes movement in the capillary zone where the water is pulled upwards first by surface tension forces and then evaporated by a thermodynamic process. Similar forces are responsible for the swelling and desiccation of the clays. The competing force which in the end draws what is left of the water to the saturated zone is, of course, the force of gravity.

The difference between one region and another in the ratio of recharge to precipitation plus imported water is the difference in the capacity and efficiency of the various "pumps". The total quantity to reach the saturated zone will be that which is left from the amount reaching the surface minus the amount retained or flowing back to the surface and to the atmosphere or leaving the region as runoff. For a direct quantitative evaluation of recharge the quantities "pumped" by each "pump" for each rain event have to be measured.Though several direct methods have been developed they require an elaborate and complicated system of monitoring and long periods of observation and even then their accuracy will still be doubtful.

Most arid and semiarid zones are regions of scarce hydrogeological data. This is due to sparse habitation as well as the random character of the climatic regime.

When the problem of recharge of groundwater is being investigated these negative aspects add to and even multiply the general variance of any hydrogeological system expressed as the variance in space and time of the hydrological and climatological parameters.

These aspects demand a rather large investment in time and money in order to collect the pertinent data needed to build the hydrological balance, especially its recharge component. In many instances, these time and money consuming conventional procedures have to be shortened in order to supply the planning engineers with a reliable hydrological evaluation as a basis for a water resource development plan. The failure to do so may either cause the postponement of a development project, which will have an immediate negative influence on the welfare of the people, or an unrealistic water development plan which may later have a negative effect on the economy of the region.

The approach suggested is a stage by stage improvement of the hydrological evaluation on the basis of knowledge acquired from other similar regions which have the same hydrogeological characteristics as the region under research. Local specific data will be progressively added while carrying out the first stages of planning and development.

It is based on the empirical knowledge that regions with similar sequences of rocks which have gone through a similar geological history and are located in similar climatic zones will have more or less the same hydrogeological character. In other words, it is suggested that a number of basic hydrogeological provinces can be defined and a procedure adopted to correlate the research area with similar hydrogeological regions. This approach is discussed more fully in Chapter 9 after a description of the principal hydrogeological provinces.

Before a preliminary quantitative evaluation of amount of recharge can be reached by correlation with similar regions, it is essential for the hydrogeologist to understand the regional water system or, in other words, to formulate the right conceptual hydrogeological model. This will help him to avoid mis-interpretation of the scarce data available; for example, to know whether the observed water tables are perched or regional, gradients true or apparent, etc.

For these purposes, first of all the characterization of the region under investigation as a hydrogeological province is needed. This characterization is possible as all the forces mentioned above, except gravity, are functions of climate, geology of the region and the character of the cover soils, (which in many instances is a function of the action of climate on the rock). Thus it can be concluded that regions with similar climatic, geological and soil conditions will also have similar ratios of recharge. Experience shows that this is the case and thus a set of climato-geological categories are proposed.

As mentioned, the soil cover is a function of both the geological and climatological conditions. In general, in more humid regions the development of a soil profile is more advanced, and vice versa. In semi-arid regions the influence of human activity on the soil cover is relatively more important than in other regions. This is due to the fact that in these regions there exists a delicate balance between the processes of soil formation and soil erosion and the intervention of man can be decisive in either a negative or positive direction. In arid and extremely arid regions the weathering processes are negligible while erosion is dominant during the short periods of the rain-storms as well as by winds during the dry periods. Due to this the higher lands in these regions are devoid of soil cover, while in the valleys depositional soil layers may be found. In many arid to semi-arid regions the deposition of eolian or pluvioeolian soils like loess is a decisive negative factor from the point of view of recharge, while eolian sands are a positive factor.

In Fig. 2.1, a schematic division into lithological categories cross-divided by iso-precipitation lines is presented. When these lithological categories (gravel, sand, sandstone, limestone and dolostone, chalk, basalt and granite) constitute the major part of a certain region they decide the hydrogeological character of the region. Such a region can then be regarded as a hydrogeological province. The most abundant provinces are the following:

(i) Alluvial fans and riverbeds

(ii) Sand and sandstone

(iii) Limestone and dolostone

(iv) Chalk

(v) Volcanic

(vi) Plutonic crystalline.

In many regions, one may find the combination of several lithologies interconnected in the subsurface. In such cases the recharge characteristics of each rock formation will have to be dealt with separately. In some regions the climatic conditions may differ in the various parts of the basins due, for example, to topographical differences. In such cases subdivisions of the area according to lithology as well as climate will have to be applied.

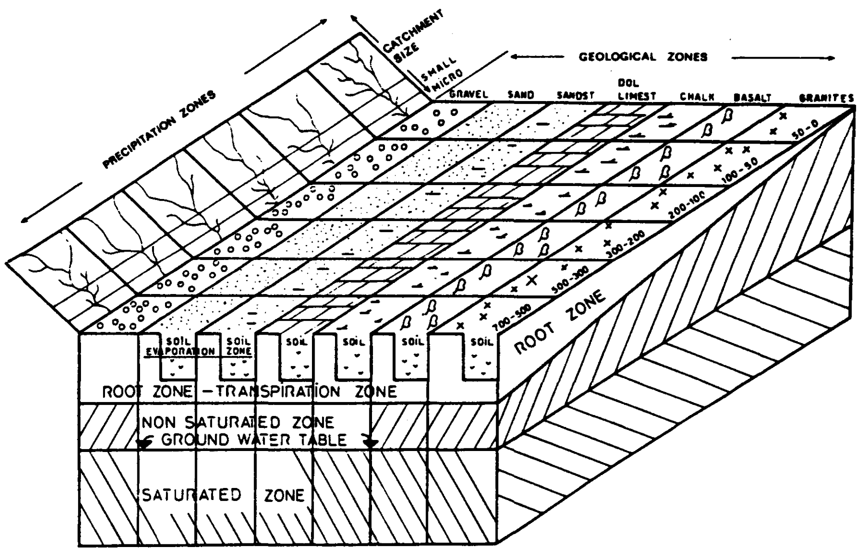


Fig. 2.1 A schematic classification of hydrogeological provinces. Based on lithological categories, subdivided by precipitation bands

3 RIVERBED AND MOUNTAIN FRONT PROVINCES

In this category two main types of sedimentary basins are included:

- riverbed basins
- mountain front basins (comprising alluvial fans, alluvial basins, piedmont plains, bajadas, etc.).

A common characteristic of these two types of basin is that recharge is by infiltration into gravel layers which usually have a rather high permeability.

Another characteristic of the recharge of these provinces is that the water that recharges the aquifers is collected from a wider area than the recharge surface. This means that the quantities which have to be considered as the potential for recharge have to be derived from runoff measurements or models extending over all the surface drainage basin.

The basic difference between the two types is that for river beds the area of supply extends parallel to the area of recharge, while in mountain front basins the supply area extends above and beyond the areal extension of the recharge area.

The combination of these two types frequently occurs in the same drainage basin when in the upper (mountainous) stretches the recharge is into a stream bed, while in the lower stretches it is along a mountain front.

3.1 Riverbed basins

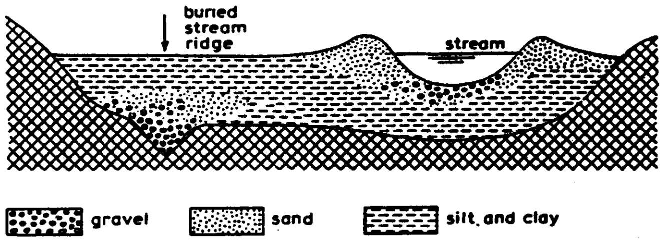
In arid and semi-arid zones most rivers are ephemeral and have a braided character when high discharge floods occur. Smaller floods can be confined to the main channel.

A major distinction can be made between meandering river deposits and braided river deposits (Fig. 3.1). The riverbed deposits are normally poorly sorted with large variations in grain size, but are predominantly coarse grained. Sediments formed within or close to river channels are much more coarse grained and permeable than those deposited on the flood plains. According to a schematic concept of depositional history, coarser grained materials should prevail in the upstream part of the plain of the river and in older, deeper layers that were deposited during earlier more vigorously erosive phases.

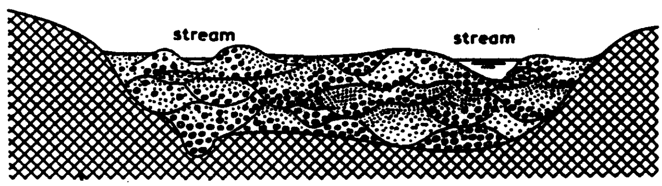
3.1.1 Physical controls on riverbed recharge

Riverbed recharge depends on the characteristics of the following (WRRC, 1980):

- streamflow
- the river channel
- the vadose zone
- the aquifer



A. meandering river.



B. Braided river.

Fig. 3.1 Cross-sections of river valleys (De Ridder, 1972)

The characteristics of streamflow which affect recharge include:

(a) Volume of flow.

(b) Type of flow: greater loss of volume is associated with flows of low discharge and long duration (i.e. winter-type flows) as opposed to flashy, shortlived summer events (Flug et al., 1980).

(c) Suspended sediment causes a lower channel permeability as it is deposited during flow recession.

Streamflow characteristics are normally easier to obtain than other data for recharge determination. The usability of a recharge model based on these data depends on whether it requires site specific calibration and, if so, whether parameters that are defined through calibration can easily be extrapolated to unstudied areas.

The channel characteristics of interest comprise:

(a) Effective channel area (length x wetted width).

(b) Permeability of the sediments.

(c) Volume of recent alluvium as this temporarily stores the infiltrated water. In general its volume tends to increase downstream.

Vadose zone characteristics can rarely be used for comparison because lithology and stratigraphy vary widely between basins. Often more useful are:

(a) Geographic position of the vadose zone.

(b) Degree of development of alluvial deposits along the stream channel. The younger deposits often have a higher permeability than the older basin fill deposits.

(c) Thickness of the vadose zone: a near surface water table will reduce recharge rates, but a very thick vadose zone (deep water table) may prohibit any infiltrated water from reaching the water table as the storage space is so large compared to the infiltrated volume. The actual situation depends on stratigraphy, permeabilities, magnitude of infiltrated volume of water and, especially, on the presence of any impeding layers acting as a barrier to downward percolation.

The aquifer characteristics are important when the water table is near enough to the source of recharge waters for a saturated continuity between the two to be possible. Important parameters are:

(a) Hydraulic conductivity as it determines the rate of transport of the recharge water. If low, the porous space fills up and reduces recharge.

(b) Specific yield because it defines the storage space available.

3.1.2 Summary of river recharge controls

WRRC (1980) summarised the major features of infiltration into riverbed deposits in the following points:

- Infiltration rates are a function of channel and flow characteristics.

- Infiltration rates increase with increases in streamflow velocity and increase with water temperature.

- Infiltration rates decrease with increases in suspended sediment inflows.

- Infiltration rates on a given reach decrease in the downstream direction because of a decrease in permeability in channel sediments and an increase in suspended sediments in the downstream direction.

- Infiltration rates will be low where the water table is very near the surface.

- The rock source area for the channel sediments influences the permeability distribution in channel sediments.

- Winter runoff events appear to have greater infiltration losses than summer runoff events.

- Loss volumes are approximately proportional to inflow only to a certain point; after that, increases in inflow volume do not appreciably increase losses. (See for example Dillon, 1980).

- Most recharge along streams draining a mountainous topography occurs from winter precipitation. Most recharge along streams draining mainly piedmont areas occurs from summer precipitation.

- In nonhomogeneous, stratified material, only a portion of streamflow loss will show up at the water table immediately following a runoff event. The rest slowly drains to the water table from the vadose zone over a period of months.

- Most stream channel recharge occurs along the main drainages of the basin because of the presence of both large volumes of highly permeable alluvium and high flow volumes.

The importance of silt on infiltration rate in riverbeds has to be emphasized. In regions where the main channel is filled by loessial deposits, Issar, Nativ and Adar (unpublished report) found that the main recharge occurs in the riverbeds of secondary streams feeding into the main channel. The investigation was carried out in the Negev highland of Israel and was aimed at estimating the infiltration rate of floodwater into the loess fill (see also Hillel and Tadmor, 1962). A network of observation holes was drilled into the colluvium overlying the bedrock and into the loess fill of the main riverbed. After a rainstorm followed by a flash flood it was found that while the holes in the colluvium showed deep infiltration, the holes in the loess remained dry in their lower part.

3.1.3 Evaluation based on water balances

Another method of evaluating recharge is by carrying out a water balance for the basin which comprises all the components of the hydrologic cycle. This needs many data. This method requires a thorough insight in the governing processes and its use is therefore often complicated. A special problem for the alluvial environment is the difficulty in separating it from the neighbouring areas. Inflow from the adjacent higher areas implies that for the assessment of the water balance often the alluvial area has to be taken as a part of a wider area of study. A thorough discussion of this method is given in Part III; only a few examples of regions with scarce data will be mentioned here.

Hillel and Tadmor (1962) described in general terms the surface and soil water regime of the central Negev (Israel). Here we are only concerned with their description of wadis. They used the water balance equation to estimate the amount of available moisture in the root zone and the deep infiltration (recharge). The annual water balance equation is:

$$r = p - (e_a + q + dW) \tag{3.1}$$

where r : recharge
p : precipitation
e_a : evapotranspiration
q : surface runoff
dW : change in storage of soil water in the root zone.

For their region of study precipitation varied between 50 and 150 mm/yr. Potential evaporation is measured to be 6 mm/d, thus yielding an average annual value of about 2000 mm. The runoff from the wadis is difficult to measure. They estimated that a rain of at least 10 mm, and an intensity of 10 mm/h (for at least short spells) is usually needed to form wadi floods. The wadis are fed primarily through runoff accumulation from neighbouring rocky slopes during floods and the amount of infiltration into the soil in the main wadi bed is estimated to be between 300 and 800 mm.

A distinction was made between gravelly and loessial wadis. In the gravelly beds the depth of water penetration is several metres. For these gravelly wadis no data were given, but Hillel and Tadmor (1962) stated that deep percolation is considerable. For loessial wadis they estimate that the deep percolation is used by the plant growth, thus the recharge to groundwater is negligible. It can thus be concluded that in similar arid regions no recharge occurs from floods in riverbeds filled by thick silt deposits (more than one metre) but recharge may occur in gravel filled riverbeds and through colluvium layers.

Abdulrazak et al. (1988) investigated the regime of recharge in a representative alluvial wadi in S.W. Saudi Arabia. They showed that each part of the wadi responded differently due to soil heterogenity. The developed regression models show that the maximum flood hydrograph depth is the most important influencing factor affecting recharge. The observation wells showed a total rise of 1.50 to 1.75 m in the groundwater table for a rainfall of about 44 mm which caused a runoff depth of 1.26 metres.

Khouri (1982) described how wadi recharge occurs in the Wadi El Miah area (Syria). Precipitation for this region varies between 70 and 150 mm. The waters of shallow aquifers are diluted by wadi recharge and fresh groundwater lenses are formed in the neighbourhood of the larger wadis. Low mineralization of the groundwater in a well (175 ppm) is an indication of the existence of such fresh water lenses.

Zaluski & Sadek (1980), however, concluded for the Wadi al Ajal in Libya that no recharge was likely to occur as the annual precipitation is very low (<25 mm/yr) and the conditions for evaporation very favourable.

3.1.4 Examples from isotope studies

Tritium, chlorine and the stable isotopes oxygen-18 and deuterium are the most widely used. Their various possibilities for deciphering the recharge rate and process are thoroughly described in Fritz and Fontes (1980) and Allison (1981). Tritium is a very effective tracer for studying the occurrence and percentage of modern recharge water in groundwater. Nearly all field studies make use of this isotope and a summary of some is presented below.

The descriptions presented here are restricted to the qualitative aspects of the use of the tracers, the recognition of recharge and its source, and semi-quantitative recharge estimations. The methods of calculation themselves are described in Chapter 11.

Quantitative tritium methods for determining the actual amount or percentage of recent recharge are described, among others, by Sukhija and Shah (1976), Colville (1984) and Yurtsever and Payne (1979). These methods are not so often used in alluvial environments as it is difficult to take representative soil moisture samples.

Gonfiantini et al. (1974) in a study of the Hodna plain in Algeria found clear evidence of recent recharge of the shallow groundwater from wadis as indicated by tritium values of 10 to 30 TU. Wells outside the wadi bed had very low tritium values, indicating that direct recharge by rainwater is relatively insignificant.

Calf (1978) in his investigation of recharge to the Namoi Valley (Australia) aquifers used tritium for the identification of recharge components. He found that the change in tritium concentration in boreholes near the river depended on several variables: the rainfall intensity distribution, the tritium levels in rainfall, the tritium levels in the river and the relative importance of various contributions to the recharge process including precipitation recharge, river recharge, recharge of river water and water from irrigation channels, and the hydraulic history of the aquifer in the past few decades. Detailed interpretation of the tritium curves was therefore not attempted, but the magnitude of the tritium levels indicated that a substantial proportion of the water was of post-nuclear times (after 1952).

Allemmoz and Olive (1980) used several methods to evaluate the recharge in a coastal plain with some minor wadis in Libya. Tritium was used to evaluate the mean residence time. This method gave good agreement with values determined by groundwater level changes. They concluded that recharge occurs mainly through wadi beds and that this recharge (1/3 of the floodwater) takes place very quickly (a few days) through large conduits of the unsaturated zone.

Wurzel (1983) described a study of the Sabi Valley alluvial plain in the extreme southeast of Zimbabwe. The underlying alluvium varies in thickness between 30-40 m near the Sabi river to 120 m near the eastern boundary of the plain. It consists of sand(coarse to fine), silt and clay. Clay horizons and lenses are present at the surface and the aquifer is confined. In the area of study a perched aquifer was found to overlie an extensive main groundwater basin. A study was made on whether any actual recharge occurs in the alluvial aquifer. Early studies (1959-68) using a piston flow model indicated a filtration velocity of 0.3-0.6 m/day. Studies of the bomb tritium peak resulted in well defined breakthrough curves that were used for the subsequent calculations. An apparent flow velocity of 0.4 m/d was obtained for the upper perched aquifers. Wurzel concluded that the river contributes only minor recharge to the main groundwater basin. He further concluded that the computed flow rate from the tritium study was in good agreement with that calculated by standard hydrogeological techniques.

3.2 Mountain front basins

Mountain front recharge is defined by Wilson, et al. (1980) as recharge occurring along a boundary of the regional aquifer system that parallels a mountainous area. For a mountain front deposit with relatively deep water table at the upper part a distinction can be made between two components (Fig. 3.2):

(i) infiltration of streamflow from the streams that drain the mountains;

(ii) subsurface inflow from the mountain mass to the basinfill sediments (also called hidden recharge by Feth, 1964).

The source of water for mountain front recharge is either direct from precipitation infiltration or indirect from streamflow infiltration. Most recharge is derived from the mountains and the importance of the size of the contributing catchment must be stressed as the recharge amount is positively correlated with the size of the catchment.

Wright (1980) stated that temporary rivers in arid zones are frequently the major source of recharge. These temporary rivers are formed in the valleys, or wadis, following intense storms in the hills which are sufficiently severe to generate surface runoff. The actual occurrence of surface flow and the subsequent onflow on a mountain front deposit depends mainly on the rain depth and the amount of evaporation.

These temporary rivers may terminate either in spreading zones on the mountain front deposit where the flood water (partly) infiltrates, or in chotts or sabkhas which are low lying areas where temporary lakes are formed. The aquifers are recharged mainly in the foothills, or piedmont zones, where the surface runoff is concentrated and where topographical conditions and soil permeability tend to be more favourable for infiltration to the saturated zone. Wright (1980) mentioned several factors whose combination favours the occurrence of recharge in piedmont zones: (i) in such areas where there is a thickness of permeable detritus comprising sand, gravel and talus (detritus fallen from a cliff face), (ii) the beds of the wadis are higher than the groundwater table,(iii) water may flow horizontally through the banks, (iv) the surface water spreads out over the ground thus accelerating the process of infiltration and subsoil saturation, (v) the finer sediments that could impede infiltration are carried to the downstream periphery of the recharge zone.

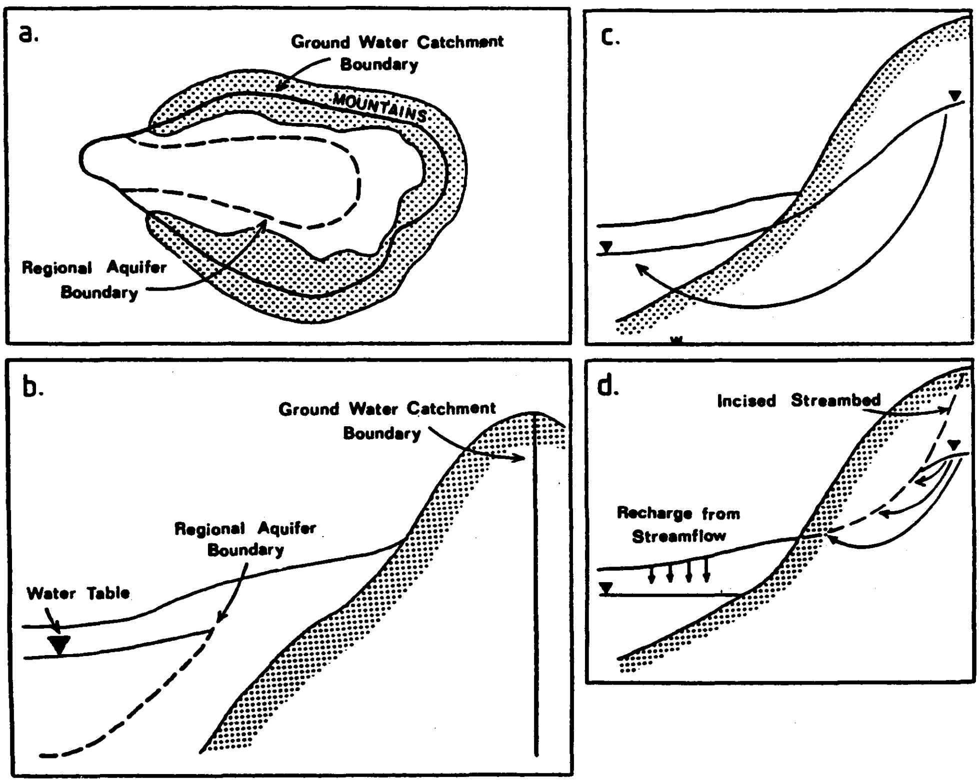


Fig. 3.2 Principles of groundwater recharge in mountain front basins. (a) Plan view of boundaries of groundwater catchment and regional aquifer (b) Cross-section of boundaries
(c) Mountain front recharge by subsurface inflow
(d) Mountain front recharge through streamflow infiltration (Wilson et al, 1980)

Alluvial fans can be divided hydrologically into three zones (De Ridder, 1972) (Fig. 3.3):

- the recharge zone
- the transmission zone

- the discharge zone

Though the recharge zone along the mountain front is of primary interest, the view that the highest infiltration capacity is confined to this upper part of the alluvial fan is often not correct. This is due to the irregular occurrence of permeable coarser deposits in the main stream channel. Huntley (1979) noted that the mountain front may have a low permeability due to the poor degree of sorting. Only a detailed study of the sedimentology of a mountain front deposit can decipher the distribution of the more pervious areas. However, as this implies a relatively large amount of fieldwork, the deposit can often be regarded as a whole and the infiltration/recharge determined by other methods which pay no attention to the exact location of the infiltration zones.

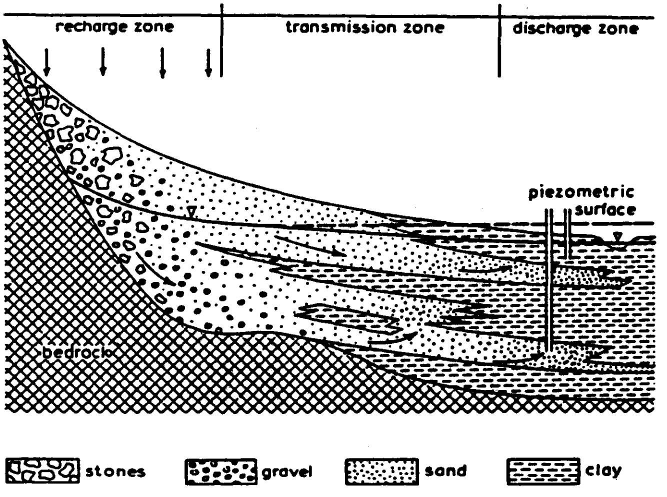


Fig. 3.3 Cross-section of an alluvial fan (De Ridder, 1970)

Wilson et al (1980) summarized the literature on mountain front recharge in the alluvial basins of the southwest U.S.A. The main points are:

(i) The amount of recharge is a function of precipitation, which again is related to elevation, relief, and orientation of the mountain.

(ii) Winter precipitation is primarily responsible for mountain front recharge.

(iii) Permeability of soils and bedrock in the mountain affects the way in which mountain front recharge occurs and also affects the rate and volume of recharge.

(iv) The topography of the mountain determines how mountain front recharge occurs. Whether or not the topography is incised enough to intersect a recharge mound determines if the recharge occurs as streamflow infiltration or as subsurface inflow.

(v) The distribution of recharge in space highly depends on the geology (stratigraphy) of the mountain front deposit.

The most important parameters are the lithology and stratigraphy of the mountain front region, because they determine the percentage of recharge from the input (the potential recharge amount). In regions where summer and winter precipitation occur winter precipitation seems to be the most effective in recharge because of lower intensity, longer duration of an individual event, lower evapotranspiration, wider areal coverage, and increased likelihood of snow.

Data related to the above mentioned conditions have thus to be collected in order that the boundary conditions as well as preliminary spatial and hydraulic parameters of the basin can be established. This will enable the formulation of a mathematical model which will be a step forward in the long term, basin wide determination of the recharge.

In general it can be said that in most cases the boundary between the mountain front and alluvial basin can be considered as a closed boundary. In special cases when the mountains are built of permeable rocks the hydraulic parameters of these rocks as well as gradient of water table have to be determined. It can be stated, however, as a general rule that in the case of a highly permeable mountain area, like karstic limestones, the quantity of flood water reaching the alluvial fan is rather small and the main hydrological research has to be concentrated on the mountain region.

3.2.1 Examples from isotope studies

Gupta & Sharma (1984) studied the recharge in the alluvial parts of the Sabarmati basin (India) using the tritium tagging method. The Sabarmati basin is an area of about 22,000 km^2 in western India. Geologically the basin is composed of crystalline rocks in the northern and northeastern parts while recent alluvial deposits cover the central and southern parts. Only the alluvial part (about 60% of the basin) was studied.

The average annual precipitation is about 800 mm, 95% of which occurs during the summer monsoon (June-September). The coefficient of variability is between 40 and 60%. The average annual potential evapotranspiration is 2700 mm.

The measurements at several sites resulted in considerable variation of the estimated fractional recharge, i.e. the ratio of net moisture transfer to the total water input (Fig. 3.4). By combining the recharge rates with geological features of the basin, the alluvial parts of the basin were divided into three zones (Fig. 3.4):

- Zone A, an area of high fractional recharge (18%) having a relatively coarse grained thin alluvium cover (medium grain size of topsoil 100 µm)
- Zone B, with intermediate fractional recharge (14%) and a medium grainsize of 70 µm
- Zone C, an area with low fractional recharge (6%) has a relatively fine grained topsoil (medium grainsize < 50 µm).

For the entire Sabarmati basin an area weighted fractional recharge of about 14.5% was found.

Estimation techniques for mountain front recharge are also discussed in Section 11.8.4.

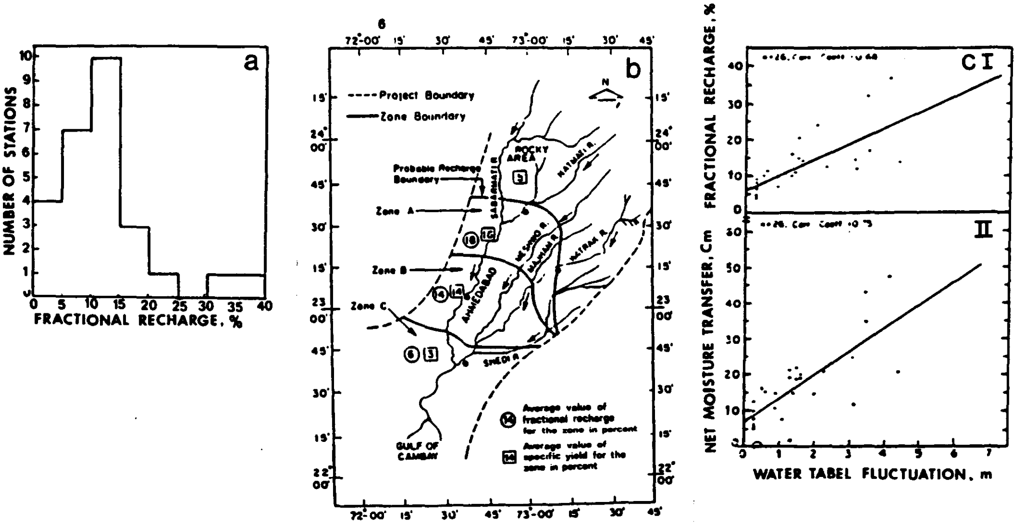


Fig. 3.4 Mountain front recharge to the Sabarmati basin, India (from Gupta & Sharma, 1984).
(a) Histogram of fractional recharge values for 27 tritium tagging stations
(b) Zones of characteristic fractional recharge values, delineated by a combination of tritium tagging experiments and conventional hydrogeology
(c) Correlation of (i) fractional recharge (ii) net moisture transfer with water table fluctation at the tritium tagging stations

4 SAND AND SANDSTONE PROVINCES

4.1 Sand provinces

These are found in regions in which aeolian agents are predominant. The source of the sand in most cases is secondary, namely, the decomposition of sandstone layers. In some regions, however, one finds sands as primary products, namely, as a product of the decomposition of crystalline rocks, the washing or blowing out of the fine, mainly feldspatic components and the transport and spreading of the sands to form dunes.

In coastal areas the source of the sands is the sea, the sand being brought ashore by currents and transported by the waves. It is then blown inland by winds. The fixation of the sands and their development into sandstones covered by soil is a function of the rate of supply as well as the climate. With a continuous supply and an arid to semi-arid climate the sands will remain loose and mobile. Their mobility and the effect of the climate will restrain perennial vegetation. The bareness of the sand dunes is an important factor in their water regime.

A different environment is formed by fluvial sands. They have widely varying grainsizes, including gravel and boulders. These deposits can either be classified as sand or alluvium; here they are considered with alluvial basins.

Sand dunes are normally composed of several successive layers which represent different depositional stages. The sand normally has a relatively uniform grain size and a varying clay and silt content.

Sand provinces are characteristic of desert plateaus where sandstones and crystalline rocks are abundant, the sand seas of the Sahara, Arabia and the Kalahari being the most famous. The porous character of the sands determines the nature of recharge which is rapid. In most cases, recharge is direct, namely, it comes directly from the rain falling on the sand area. The flow through the unsaturated zone in most cases can be considered as a vertical piston type flow.

The amount of water infiltrating into the surface of the sand dunes is referred to as potential recharge (Verhagen et al., 1979). The recharge is normally supposed to be only direct recharge, but this is probably confined to coarse grained sand dunes. Localised recharge (as defined in Section 1.1.2) is probably common as the infiltration water often first flows as runoff over a (small) distance (Gat, 1984).

For the recharge rate in sand dunes special attention must be given to the presence of consolidated horizons which may block the downward movement of the soil moisture (e.g. hardpans, "kurkar",etc.).

The presence of moisture in the sand layers depends a great deal on the climate and on the size distribution of the sand.

Dincer (1980) stated that the mean grain size of dune sands often varies between 0.15 and 0.4 mm. Sand dunes with coarser grains let the rainwater infiltrate to considerable depth owing to their low field capacity. While no moisture could be found in sand dunes with a mean grain diameter of 0.2 mm in Saudi Arabia (except after rainy days), sand with a similar grain size in the northern Kalahari desert had ca. 7% moisture by volume, due to the much higher precipitation (500 mm compared to 70 mm at the Saudi Arabian site).

4.1.1 Examples

The examples of recharge studies given in this section on sand provinces have been divided into precipitation classes to distinguish different recharge classes. As the natural processes of recharge are influenced by the total quantity of precipitation as well as by the regime of the rain storms, the recharge values are directly related to the precipitation class to which the specific area belongs. Therefore a distinction into different recharge classes can be made, stressing the fact that even under very low precipitation regimes, recharge can occur. The recharge is further a non-linear function of the precipitation. Three groups are introduced and illustrated by some characteristic studies.

The usefulness of this division into different recharge classes is limited by the fact that the literature does not cover all the precipitation classes. Most studies are confined to the arid zones (e.g. < 100 mm annual rain) or to the semi-arid/sub-humid zones (> 700 mm). It must also be admitted at this stage that the average annual rainfall is not an optimal indicator of climatic conditions in very arid regions as the variance in this figure is often very high (Lloyd, et al., 1987). In combination with the variance in precipitation and with the evaporation a more accurate division might be possible, but these data are not always available.The different zones as indicated on Fig. 2.1 have been divided into three groups:

(i) 0-300 mm

(ii) 300-500 mm

(iii) >500 mm.

(i) 0-300 mm The most arid regions have been studied in Saudi Arabia (Sonntag et al., 1980; Dincer, 1980; Caro & Eagleson, 1981), Tunisia and in New Mexico (Phillips et al., 1984). In this class soil development is nearly absent. For the Saudi Arabian site the annual rainfall was 50-100 mm. Dincer et al. (1974) had calculated a recharge rate of 2.3 mm/yr.

Vachaud et al. (1981) and Vauclin & Vachaud (1981) studied the infiltration and deep percolation at a site near Gabes (Tunisia). The average annual precipitation was 183 mm, with extremes varying between 36 and 532 mm. The potential evaporation is estimated at between 1420 and 2000 mm. The soil profile consists of a sandy layer 90 cm thick, underlain by a gypsum crust between 90 and 110 cm.

Using a rain simulator, they discharged 173 mm at an intensity of 19 mm/h (9 hrs of constant wetting) onto the soil surface. The deep percolation at 220 cm depth amounted to 131 mm (76% of the artificial rainfall). This relatively high value of infiltration is most probably due to the particular constant artificial rainfall intensity. It can nevertheless be concluded that rain events of high intensity and duration of a few hours (which are known to occur from time to time in these arid regions) will recharge the sands at the above mentioned rate.

Caro and Eagelson (1981) studied recharge in sand dunes in Saudi Arabia. For an area with an average annual precipitation of 129.8 mm, they found a recharge of about 14 mm, i.e. about 10%, while an area with an average precipitation of 79.5 mm yielded a recharge of about 6 mm, i.e. about 7.6%.

In New Mexico, which has an annual rainfall of 200 mm and an estimated potential evapotranspiration of 1800 mm, Phillips et al. (1984), using bomb chlorine-36 as a tracer, calculated a net infiltration rate of 2.5 mm/yr,which is 1.2% of the annual precipitation. On the other hand, in the same country, Stephens and Knowlton (1986) investigated the infiltration into alluvial sands by measuring the soil water movement using tensiometers and neutron probes. Recharge was calculated from Darcy's equation using in situ pressure head data below the root zone, calculated hydraulic gradient and in situ hydraulic conductivity. Recharge was also calculated as being equal to the unsaturated hydraulic conductivity corresponding to in situ water content. Both methods gave the annual recharge rate of about 3.7 cm/year which is about 20% of the mean annual precipitation for that period of 1982-1984. The dissimilarity between these two results may be due to the different methods. The latter being a more direct method seems also to be closer to reality.

A study of the recharge rate in the Thar desert in India, with an annual rainfall of 150-300 mm, yielded values of 6-14% in the dune sands (Sharma & Gupta, 1985).

<u>(ii) 300-500 mm</u> For this climatic zone much literature exists as this group includes the intensively studied Kalahari desert.

The southern edge of the Kalahari within the catchment of the ephemeral Gamagara river receives an annual rainfall of 330 mm (Verhagen et al., 1979). From the moisture balance, using tritium profiles, they found strongly different values of infiltration in the sands (3.5-26.6% of annual rainfall).

The Central Kweneng area in the Kalahari was studied by Foster et al. (1982) through the measurement of the hydrophysical properties of the sands as well as the Cl, Na and tritium profiles. The annual rainfall here was 450 mm, with a coefficient of variation of about 65% (300mm). Mean daily evapotranspiration reaches values up to 8 mm. Downward movement through the sands was estimated to be certainly less than 5 mm/yr and probably less than 1 mm/yr. In general, it can be concluded that in regions with a rainfall regime similar to that of the Kweneng area, no diffuse groundwater recharge will actively occur through a sand cover of more than 4 meters depth. The water which infiltrates to this depth will be temporarily stored and used by the vegetation during the dry seasons.

Allison & Hughes (1983) studied (with the aid of environmental tritium and radiocarbon) an area with a mean annual rainfall of 335 mm. The soils consist of sands, sandy loams and sandy clays. A comparison was made between vegetated areas (mallee) and stripped areas. Only for the vegetated areas did they give a recharge value: 3-4 mm/yr.

For this recharge class the effect of macropores becomes important as the vegetated area is more abundant under this precipitation regime. Macropores are often formed along existing roots or by the empty channels of dead roots (Allison & Hughes, 1983).

Edmunds & Walton (1980) studied recharge in the dune sands in the extreme south of Cyprus with a mean annual rainfall of 420 mm. With the use of different methods a recharge rate of about 50 mm/yr was determined. The profiles of electrical conductivity and chloride were calibrated with the precipitation data to obtain a recharge graph over the time of record (Fig. 4.1).

Cable (1980) studied infiltration into sandy loams and gravelly sandy loams by measuring moisture profiles in the soils with the aid of a neutron probe. He noticed the difference the rainy season made on the amount of infiltration, similar to the effect found in alluvial environments. The area of study at an experimental range south of Tucson (Arizona, USA) receives an average annual rainfall between 270 and 430 mm. About 56% falls from July through September, and most of the remainder from December through April. The two rainy periods are thus separated by dry periods, of which the May-June drought is particularly severe. Winter storms are typically extensive, of long duration and low intensity. Summer storms are localized, intensive thunderstorms of short duration.

In the same way as described for recharge in alluvial basins in Section 3.2, the winter rains are the most effective for recharge. Although more precipitation usually occurred in summer than in winter, moisture from winter rains penetrated deeper and lasted longer. The relatively high effectiveness of winter moisture, especially on bare soil, is attributed to:

- typically, the intensities for winter precipitation do not exceed the infiltration capacity of the soil, whereas the summer storms normally do;

- the small droplets characteristic of winter storms fall at lower velocities and result in less soil splash and sealing of the surface; and
- evapotranspiration demands are much lower in winter than in summer.

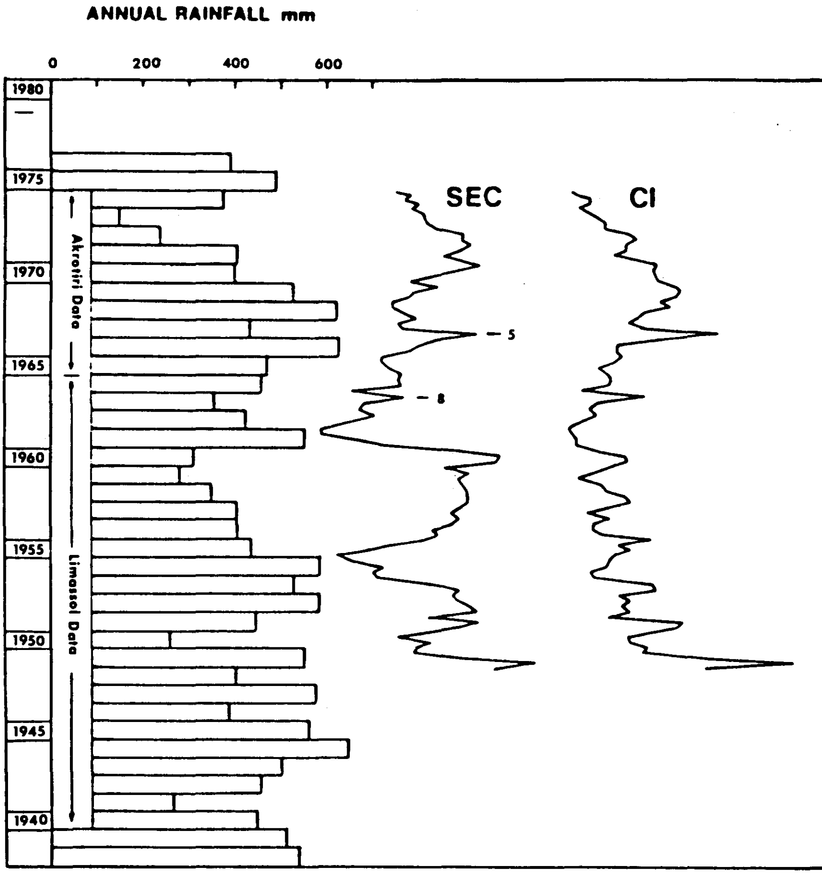


Fig. 4.1 Time profiles of rainfall, conductivity and chloride for Akotiri, Cyprus. For period 1938-76, with location of specific electical conductance (SEC) and Cl profiles derived from position of Tritium peak (Edmunds & Walton, 1980)

Ground cover of plants and litter greatly increases the effectiveness of the summer precipitation, but increases the effectiveness of the cool season precipitation only moderately.

In his conclusions, Cable further stated that bare soil recharged almost as well as vegetated soil in winter, but summer recharge of bare soil was only one third that of vegetated soil, and there was little infiltration below 25 cm.

Well vegetated soil recharged about as well in summer as in winter. Perennial grass cover increased the soil water supply in summer by increasing infiltration, decreasing runoff, and slowing evaporation. There was rarely any recharge from one rainy season to the next because all available water usually transpired or evaporated by the end of each depletion period. Available water was reduced to between 1 and 2% by volume at all depths by the end of most depletion periods.

<u>(iii) >500 mm</u> The least arid class includes the studies which are on the edge of what is normally considered the semi-arid zone (i.e. 700 mm, Lloyd, 1980).

Allison & Hughes (1974) made a study of recharge to an unconfined aquifer through soils composed of clay, loam and sand by studying tritium profiles. The mean annual rainfall is about 750 mm, most of which falls in winter. The land cover at all sites is improved pasture. The mean annual recharge varied between 40 and 140 mm, dependent on the soil type.

Three sites in the Gambier plain (approx. 140.8°E, 37.8°S), Australia, were chosen for their studies:

Site A has a soil profile consisting of 0.5 m of sand overlying heavy clay which grades to highly calcareous clay at 4 m. These overlie a layer of recemented limestone at 4.7 m above a sandy aquifer. The water table varies over the range 4.5-5.5 m depth. Using different models the recharge rate was found to be 40-50 mm/yr.

Site B was situated on sandy loam overlying sandy clay. No further details were given of the profile. Here a recharge rate was found of 80 mm/yr.

Site C has a profile of 0.6 m of sandy loam overlying 6 m of clay interspersed with occasional thin sandy lenses. The soil has a high clay content, but is highly structured and allows for relatively rapid downward movement of water. A sandy horizon at 7 m is underlain by a thin layer of carbonate cemented sand and white carbonate-rich clay. The water table is at about 15 m depth. The recharge rate was calculated to be 130-140 mm/yr.

The large differences in recharge rate for the different soil type environments can be partly explained by the differences in available soil moisture storage. This amounts to 200 mm for site A and 100 mm for site C.

The study of the "Gnangara Mound" in the Swan Coastal Plain (Western Australia) by Sharma & Hughes (1985) using chloride, deuterium and oxygen-18 profiles, showed that on deep sandy sites (>20 m), current recharge under the mature dense pine plantation was negligible, while under an adjacent Banksia woodland site it was about 11% of the average annual rainfall (775 mm/yr). However on a relatively shallow sand (<10 m), recharge was about 7% under a pine plantation and about 25% under the Banksia woodland.

A long term average of recharge in this region is suggested by Sharma (1988) to be 15% of average yearly precipitation (775 mm/yr); over 50% of this recharge occurs through preferred pathways.

In the Swan Coastal Plain, Carbon et al. (1982) used neutron scattering techniques to measure soil moisture in a study which showed that the replacement of native forests with perennial pastures would have an effect on deep drainage. The replacement pastures would lead to a significant increase in deep drainage. In the Sparwood Dune, recharge was above 30% of annual precipitation (800-900 mm). In his study of the adjacent Bassendean Dune System, Balleau (1973) had found that the net recharge through native woodlands was more than 30% of annual rainfall.

4.1.2 General conclusions

From these data for the different classes it becomes clear that recharge rate is dependent on, among other factors, soil type, vegetation cover, precipitation distribution over the year and other meteorological variables. It does not seem possible to give an absolute percentage interval for the recharge as part of the annual precipitation of an area. It is, however, possible to give a summary of the characteristics of the recharge to sand environments.

Such a summary was presented by Issar et al. (1985) on the basis of results of various investigations reported from the sand dune areas of the central and southern coastal plain of Israel. In the central part, where the average annual precipitation amounts to 500 mm (winter rains), empirical methods were applied (see also Section 11.3). The formula arrived at is $r = 0.81$ $(p-94)$ where r is recharge and p is precipitation, both in mm/yr. This gives recharge of about 65% of the average annual precipitation. In the southern coastal plain, where the average annual precipitation amounts to 200 mm, the annual outflow to the sea was calculated by using Darcy's flow equation and field data regarding the hydrological parameters. The average annual recharge ranged between 55 and 60% of annual average precipitation.

In the inland dunes, where the average annual precipitation amounts to 100 mm, neutron probe observations in the sand dunes showed that 60% inflow of water was obtained after rainstorms of more than 5 mm, yet no perched water table was observed in areas where the riverbed cuts the dunes and their bases can be observed to be overlying semipermeable loess. The reason for this is most probably that these dunes were reforested and the trees transpire the water which could have formed a perched groundwater table.

The more general conclusions are the following:

(i) Direct evaporation and transpiration by annual vegetation assuming a constant supply of water (for example, by irrigation) amounts to 30% of the water applied; thus, under favourable conditions, for example, heavy rainstorms, about 70% of the annual precipitation may infiltrate below the root zone of the vegetation.

(ii) In arid to semi-arid regions less water will reach the groundwater due, first of all, to the fact that some rainstorms are below the minimum amount or threshold. This "minimum amount" will depend on the type of dune vegetation and on the storage coefficient of the sand. In general, it can be estimated that isolated rainstorms of less than 5 mm, namely those which saturate the sand (storage coefficient = 10%) to less than 50 mm depth, may be lost for groundwater replenishment in a region where annual vegetation exists. As such storms account for about 40% of the precipitation in the central Negev area, one can estimate a priori that groundwater replenishment will not be more than about 60% of the precipitation.

(iii) In regions with deep rooted perennial vegetation, groundwater replenishment will be a function of the amount transpired by this vegetation. It can be estimated that in areas with perennial vegetation, with a root zone depth of 1.5 m, a minimum accumulation of 150 mm of rain water is needed for a certain percentage to permeate through the root zone and to replenish the groundwater.

(iv) Taking all these considerations into account, the general equation for infiltration, suggested for an area like the coastal dunes, is as follows:

$$r = 0.4\ (p - t_r) \qquad 4.1$$

where r is recharge to groundwater, p is annual average precipitation, t_r is annual transpiration from deep rooted vegetation, all in mm. The factor 0.4 accounts for the direct evaporation and transpiration from annual vegetation as explained above, and in areas where annual precipitation is around 100 mm and which were reforested recharge to groundwater will be nil.

For recharge estimation the salinity and isotopic composition of the groundwater can also be used as an indicator. For the recharge determination which takes into account the different processes including infiltration and subsequent percolation, use can be made of flow models. (See also Sections 11.6.2 and 16.5.3-16.5.4.)

4.2 Sandstone provinces

These provinces are abundant along the margins of the ancient continents or shields. They are a result of the continuous deposition of sands through geological ages, or even eras, in regional sedimentary basins. The sedimentary basins were mainly continental terrestrial, and the sands were deposited as aeolian dunes which were later consolidated due to lithostatic pressure and precipitates. Some of the sandstone was deposited in lacustrine or even shallow marine basins. In these cases, the clay content increased to a sufficient quantity to form continuous impermeable layers. Thus, the more continental and more arid the paleo-environments were, the more continuous and homogenous the aquifer system is.

On a regional scale, one finds thick, continuous, phreatic aquifers in the region bordering the ancient shields while towards the ancient seas or lakes the sandstone sequence subdivides into secondary aquifers. The aquifers become confined in this direction because the general regional dip of the layers is from the ancient continent towards the sea or lake.

The recharge thus takes place in the phreatic zones on the margin of the crystalline shields. Local rain provides the direct recharge while indirect recharge comes from the rivers draining the shields.

The big thickness of the layers of sandstones in the regional aquifers enables a very large amount of storage. In arid zones the annual recharge may be negligible in relation to these amounts. Thus development of groundwater in certain regions is based more on subsurface flow from adjacent regions than on vertical recharge.

In the confined regions the flow of water is very slow and residence time of the water in the aquifers exceeds thousands of years. In these cases the question of natural recharge has no practical significance from the point of view of safe yield or modelling. As the storage capacity of such basins is tremendous, long residence time is the main characteristic and thus paleo-climatic effects are considerable, if not decisive.

The recharge to sandstone aquifers can be divided into two classes:

(i) recharge directly into the outcrops of the sandstone formation,

(ii) recharge through overlying deposits, e.g sand dunes.

The second class strictly belongs to the preceding chapter, but in the literature often no distinction is made between these two classes. The infiltration of precipitation water into overlying deposits such as alluvium gravels and sand dunes is often considered as delayed input into the underlying formations. The infiltration and partly also the subsequent recharge is, however, largely determined by the overlying deposit and the mechanism is thus described in the corresponding chapter. In this way, recharge into the sandstone formations may occur in dry seasons due to the storage capacity of the overlying deposits.

The recharge to such large scale aquifers is often confined to certain periods in the geological history of the basin. The higher historical recharge rates are often due to climatological conditions different from the present conditions. These conditions can often be identified by the chemical and, especially, the isotope composition of the (fossil) water.

The recharge mechanism is a function of the geological history of the basins. The most typical are those along the margins of the ancient shields. In these the recharge is mainly along the zone between the basin and the shield.

For the sandstone provinces, most studies describe only the occurrence and processes of recharge and give the actual amount of recharge only as volumes, e.g. in millions of m^3. We give these values in the present publication without converting them to percentage of precipitation. The reason for this is that conditions in one region differ markedly from another. Moreover, in most regions, the area of net outcrops has not been given, nor whether the recharge is direct or non-direct or if water is flowing from adjacent areas.

4.2.1 The Great Artesian Basin, Australia

The Great Artesian Basin (GAB) has been extensively studied since the seventies and especially the groundwater flow pattern has been investigated (Habermehl, 1980, 1983; Forkasiewicz, 1982, Calf & Habermehl, 1983). A detailed account of the hydrogeology of the GAB is given in AWRC "Groundwater resources of Australia" (1975).

The GAB is a confined groundwater basin comprising aquifers of continental quartzose sandstones and confining beds of siltstone, mudstone and marine argillaceous sediments. The Mesozoic sedimentary sequence reaches a maximum total thickness of about 3000 m in the central part of the GAB. It occupies an area of 1.7×10^6 km^2 and is located in the arid and semi-arid zones of Australia.

The annual rainfall on the basin varies from 100 mm in the southwest (Simpson desert) to 600 mm in the extreme eastern part. The potential evaporation varies between 1800 mm in the northeast to 3300 mm in the southwest (Forkasiewicz, 1982).

Recharge areas in the GAB have been interpreted from the piezometric contours. Delineation of the recharge areas has been validated by different environmental isotope and hydrochemical studies (Airey et al., 1983). The outcrop areas of the aquifers provide the major recharge to the aquifer units and occur mainly along the eastern margins of the Basin (Habermehl, 1983). Along the western margins, additional smaller recharge amounts are believed to infiltrate in the outcrops overlain by sandy sediments.

Estimates of recharge were made for different parts of the GAB grouped according to the State to which it belongs, but only for the state of Queensland is a number given. For this region, which comprises a great part of the eastern recharge area, the amount of natural spring outflow and underflow to other states was estimated to reach 295 000 m³/d (0.1x10⁹ m³ per year) (AWRC, 1975). This was thought to be far less than the potential recharge available in an estimated area of 75 000 km² receiving an annual rainfall of 635 mm/yr (i.e. about 50x10⁹ m3 per year).

Forkasiewicz (1982) mentions a recent estimate (1980) of annual recharge of 1.1x10⁹ m³.

4.2.2 Eastern Sahara

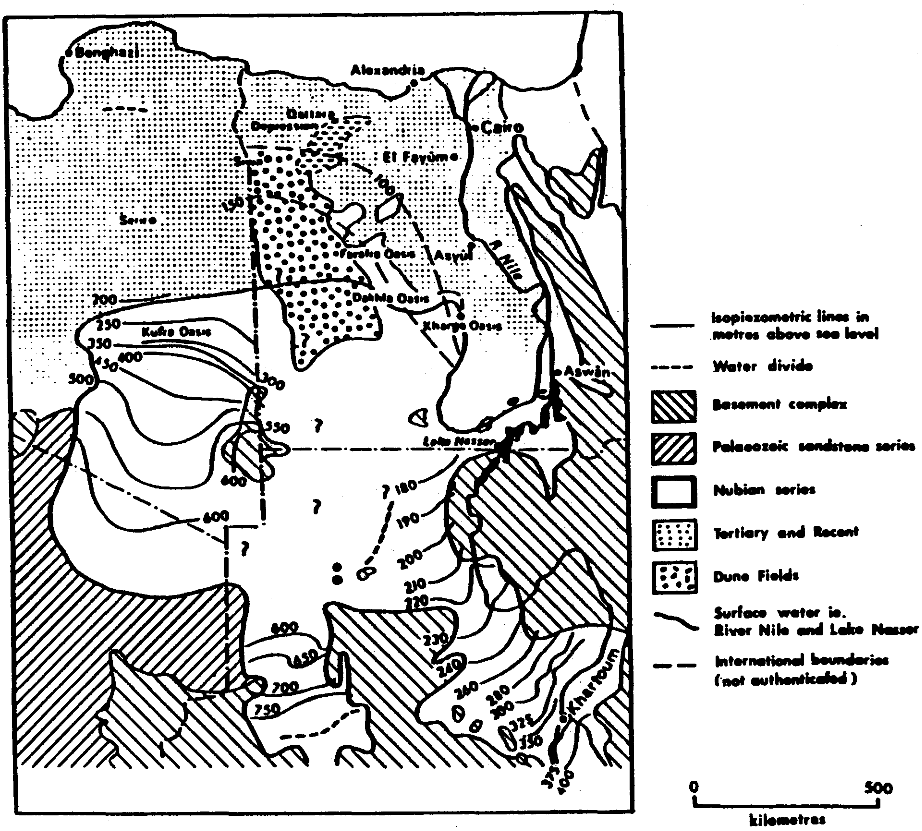


Fig. 4.2 Hydrogeology of the Nubian basin (Forkasiewicz, 1982)

Another well studied sedimentary basin is the Nubian Aquifer System of North Africa, including Sinai and the Negev

(Fig. 4.2). It has a surface area of $1.8x10^6$ km^2. The layers of sedimentary deposits, with an average thickness of 200 to 2000 m are made up of sands, sandstones, conglomerates and clays. The Nubian basin comprises a semi-closed multi-layered aquifer system, unconfined at the southern periphery of the basin, semi-confined to confined in the centre and totally confined near the discharge area in the North (Forkasiewicz, 1982).

The hydrogeology of this basin has been studied thoroughly and good summaries are to be found in Himida (1970), Burdon (1977), Forkasiewicz (1982), Shata (1982), Sonntag (1984) and Heindl & Hollander (1984).

The contemporary recharge to the aquifer is considered by nearly all investigators to be negligible (e.g. Sonntag, 1984), but still some data do exist on the water balance.

Heindl and Hollander (1984) mention that only in the Gezira area near Khartoum is infiltration from the Nile river to the groundwater proven, but it is not clear whether this contributes to the main groundwater body of the Nubian system. Gischler (1976) mentioned a loss of almost 10^9 m^3/yr of the Nile between the union of the White Nile, Blue Nile and Atbara rivers in Sudan and Aswan in Egypt and states that of this amount 50% (i.e. $0.5x10^9$ m^3/yr) may recharge the large Saharan aquifers of Nubian sandstone in those sections where the river has cut its valley into these formations. Gischler also stated that recharge takes place into the banks of Lake Nasser, created in 1964.

Today groundwater recharge by precipitation is limited to very few areas. The highest recharge may be expected in the surroundings of Haruj and Tibesti mountains (Wright, et al.,1982), though annual precipitation is often below 25 mm. Wright et al. (1982), in a study of the hydrogeology of the Kufra and Sirte basins in Libya, stated that some recharge by direct precipitation is possible. Favourable circumstances include the flat terrain largely devoid of vegetation, high infiltration capacity of the loose surface sands and gravels, and the high intensity of storm rainfall which falls in the winter months when evapotranspiration rates are low. Pachur & Braun (1980) with theoretical calculations obtained a recharge of either 50 or 25% from a typical arid zone precipitation of ca. 100 mm, the alternative value depending on whether the rainfall was summer or winter.

Other recharge areas may be the north of Sudan and some places east of the Nile river in Egypt. Heindl & Hollander (1984) state that quantitative estimates of recharge have to rely on general soil moisture balances, but should be subsequently refined by fieldwork.

Forkasiewicz (1982) when describing in more detail the several parts of the Nubian basin, estimated the recharge from the Tibesti mountains (i.e., indirect recharge) to be in the order of "some million m^3/yr". From the literature on the Egyptian part (especially the New Valley Project), Forkasiewicz (1982) summarizes four different estimates (0.59, 0.81, 1.22, 4.0×10^9 m^3/year), and concludes that the most probable value will be about 1×10^9 m^3/yr (ca. 31 m^3/s).

Wright et al. (1982) stated that although no data on wadi flows in the Tibesti region are available and rainfall records are available for only one site (Zouar), some general estimates of possible runoff and recharge are worth making. Taking the area above 800 m elevation and assuming an average value of rainfall of 37 mm, which is conservative, a 10-20% order of recharge would be equivalent to between 74 and 148×10^6 m^3 for the Kufra basin and $106\text{-}212 \times 10^6$ m^3 for the piedmont leading to the Sirte basin. The recharge compares quite closely with the estimated outflow from the Kufra basin of between 70 and 160×10^6 m^3. However, the aquifer cannot be in equilibrium with this marginal recharge since otherwise, in its phreatic condition, the age of the water at Kufra would be expected to be within the current climatic period (i.e. ca. 5000 years), which it is not. Since the rainfall of this region is extremely low it is assumed that the recharge to the basin is in marginal areas from surface runoff and from these marginal areas the younger waters spread and mix with the older water.

4.2.3 Western Sahara

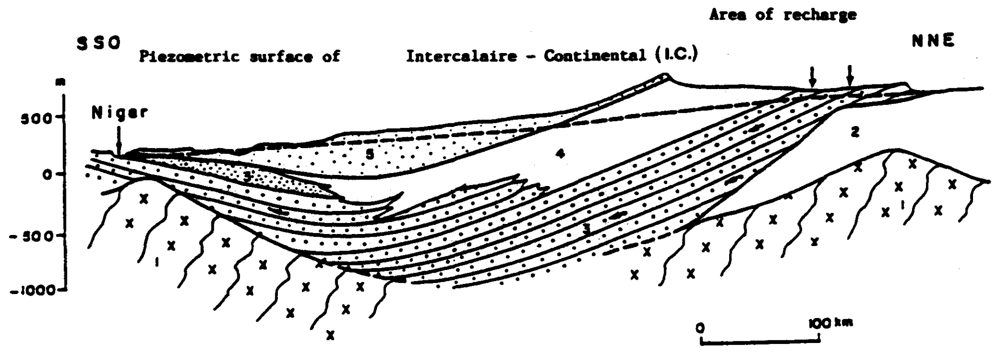
Gischler (1976) described the aquifer of the Continental Intercalaire (C.I.) underlying the western Sahara. It extends over 600 000 km^2 with an average thickness of 250-600 m. The effective porosity is 20% and the transmissivity is 0.001-0.050 m^2/s. Recharge takes place along the margins through infiltration of runoff water from the Atlas mountains and others. This amounts to about 8.5 m^3/s recharge.

The aquifer of the Complex Terminal (east of the C.I.) as described by Gischler (1976) consists of more or less ferruginous sandstones with phreatic, confined and artesian aquifers owing to clay lenses concentrated in some parts of the profile. The aquifer covers an area of 350 000 km^2 and has an average thickness of 100-400 m. Recharge occurs along the marginal mountainous zones and amounts to 18.5 m^3/s.

Margat (1982) describes the hydrogeology of the sedimentary basin underlying Mali, Niger and Nigeria (Fig. 4.3). The area has an arid climate with an average rainfall of 0-50 mm/yr.

The multi-layer aquifer consists of sand, sandstone and argillaceous sand. It is unconfined at the edge and becomes confined towards the centre and in the South. The average thickness is 240-300 m in Nigeria and 500 m in Niger.

Recharge to the basin occurs in Niger through outcrops into the unconfined zone and is estimated to be about 850×10^6 m^3/yr (i.e. 20 m^3/s).



1. Basement - metamorphics.
2. Argilaceous substratum of I.C.
3. Sands of high permeability. I.C.
4. Semi-permeable cover of I.C. and sandstones of mid to upper Cretaceous.
5. "Continental Terminal" Tertiary aquifers.

Fig. 4.3 Hydrogeological section of the Niger basin (Margat, 1982)

4.2.4 Saudi Arabia

Otkun (1971,1972) summarized the hydrogeology of both the Paleozoic and the Mesozoic sandstone aquifers of Saudi Arabia (Fig. 4.4).

Within the Paleozoic sandstone sequence, only the Saq, Tabuk and Wajid formations are important as aquifers. For the Saq, the recharge looks questionable as measurements after heavy rains in November 1967 showed no rise in the water levels in the wells located on the outcrop. For the Wajid aquifer in southern Saudi Arabia the data collected did not allow for an accurate estimation of the recharge, but an evaluation of these data showed that annual recharge is ca. $110x10^6$ m^3, which includes $9x10^6$ m^3 of direct recharge, $71x10^6$ m^3 from runoff and $30x10^6$ m^3 from underflow (Otkun, 1971).

The Mesozoic sandstone formations in Saudi Arabia can be divided into two formations: the Minjur and the Wasia formation (Otkun, 1972).

The Minjur formation comprises sequences of sandstone and shale limited by limestone at top and bottom. It has an outcrop over a distance of more than 1000 km. Like all sedimentary formations in Saudi Arabia, it dips generally eastward, with an average slope of 15 m/km. Recharge is somewhat different in the north from that in the south. In the northern region, the Minjur sandstone outcrop covers large areas which are partly overlain by sand dunes. The recharge occurs either through direct rainfall or runoff from the precipitation which falls on other formations and flows towards the Minjur outcrop. Runoff occurs only over a short distance, and the water accumulates in certain depressions.The existence of relatively impermeable soil layers at the bottom of these depressions creates adverse conditions for infiltration. Otkun (1972) came to the conclusion that only 1.5% of the rainfall is infiltrating underground during average and wet years. From the percentage of infiltration and the average precipitation (ca. 75 mm/yr), it is estimated the the recharge is ca. 1500 $m^3/yr/km^2$ (or 1.5 mm/yr). Small amounts of recharge are thought to occur also in southern Saudi Arabia under a regime of lower rainfall on a limited extent of outcrop. This recharge probably occurs from the groundwater stored in the alluvium covering the outcrops (wadi sediments).

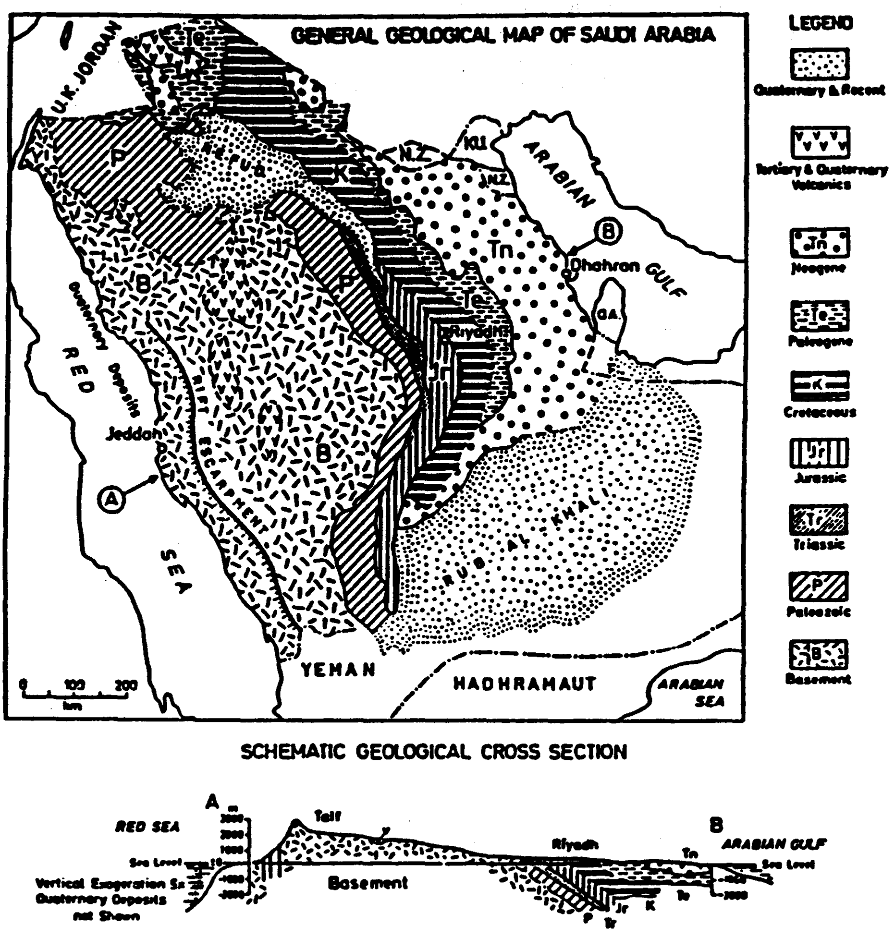


Fig. 4.4 General geological map of the Arabian peninsula

The Wasia formation is also composed of sandstone, shale and limestone. It is one of the most prolific aquifers of Saudi Arabia. The outcrop is about 1450 km long, in the form of a narrow belt. The estimates of recharge to the aquifer made by different companies are very different. ARAMCO (1959) considered (a) the percolation from rainfall on the outcrop

(b) inflow from the wadi system and (c) leakage from the aquicludes located above and below the Wasia, and concluded that the annual recharge is in the order of $2.8x10^6$ m^3. SOGREAH (1970-71) considered (a) direct rainfall, (b) runoff from the parts of the catchment extending over the neighbouring formations, (c) that part of the rain water will evaporate immediately on the hot ground and (d) part will percolate underground (depending on the varying degrees of runoff and surface storage). They concluded that the recharge is roughly 3-5 mm per year on the average. They also estimated that about 1% of heavy rainfall on a catchment reaches the floodplain area and joins the Wasia aquifer. Under these conditions, the estimated recharge will be 3000 $m^3/km^2/yr$. The data from both estimates are difficult to compare as SOGREAH only considered a part of the Wasia area.

5 LIMESTONE AND DOLOSTONE PROVINCES

5.1 Introduction

Limestones and dolostones are sediments of a rather shallow warm marine environment. Thus, one finds the regions built of these aquiferous rocks extending in a very wide belt along the ancient shores and shelves of the seas surrounding the ancient continents. The more or less constant geological conditions which prevailed during long geological periods allowed the formation of very thick and rather homogenous sections which caused the formation of aquifers extending over very wide areas to great depths.

Another important feature of these provinces is that they extend over the global belts which went through folding movements. This caused them to form mountain chains. These, like the mountains of Anatolia, Iran, Atlas and Lebanon, get higher rates of precipitation which recharge the very thick aquifers and flow out along the arid foothills as very big springs.

In most cases, limestones and dolostones form secondary aquifers, since their permeability is a function of secondary solution processes by acidized rainwater. The extreme cases of such processes are the karstic phenomena which bring the permeability of the rock and, as a consequence, the recharge to maximum. In such regions, surface runoff is nil, and in areas devoid of vegetation because of the lack of soil cover, the recharge accounts for most of the amount of precipitation in average rainstorms.

In arid and semi-arid regions the karst, although important, is not developed to its extreme under existing climatic conditions, though this may change towards the more mountainous regions which form topographic humid zones.

The vegetation cover over carbonate terrains is also important. Once established, the vegetation maintains the soil cover and humus thus helping in the formation of carbonic acid which promotes the solution processes. On the other hand, vegetation increases evapotranspiration. In general, it can be stated that vegetation is a negative factor in the recharge process, especially when reforestation is undertaken. The quantity of moisture transpired by the vegetation is a function of the type, density and climate and in each region this factor has to be investigated separately.

Though in laboratory tests dolomite is less soluble in carbonic acid than calcite, in nature solution phenomena in dolostone provinces are of the same order of magnitude as limestone provinces.

As part of the "Guide to the hydrology of carbonate rocks" (UNESCO, 1984), Zebidi briefly reviewed the infiltration (recharge) characteristics of limestone terrains. Though most of his review is based on limestone hydrology in humid regions, the general characteristics of infiltration described by Zebidi also apply to the arid and semi-arid zones. Two types of infiltration are identified:

(i) a distributed type of infiltration which is of importance for the greater number of limestone outcrops.

(ii) infiltration of a localized nature in the beds of certain streams, or in an extreme case where a whole river is engulfed in a ponor (swallow hole).

Although the infiltration depends on the annual amount of precipitation, its seasonal distribution, and above all on its intensities, only rarely is the infiltration capacity of a karst formation less than the intensity of the precipitation. The infiltration is also influenced by the topography. A young topography favours a distributed type of infiltration compared with a matured topography where there are possibilities of benefiting from larger localized infiltration.

Zebidi summarizes several studies of determinations of infiltration, often expressed as percentage of annual rainfall. From these studies it is clear that large variations occur in the infiltration rates. Zebidi states that it is advisable to determine an average infiltration rate which filters out the large fluctuations. Yet this is not the simple arithmetic mean of seasonal or annual recharge figures. Therefore it is better to use a water balance calculation over as long a time period as possible, where the recharge figures are one component among others. The following examples are based on Zebidi (UNESCO, 1984).

5.2 Examples

5.2.1 Tunisia

Schoeller (1948) studied a limestone structure covering an area of 5.8 km^2 in Tunisia. A great variation in the rate of infiltration (30 to 90%) was found, a variation which was a function not only of the amount of rain (348-676 mm/yr), but also of the intensity of the precipitation and the state of the groundwater reserves at the end of the dry season.

Large variations in infiltration rates were also found by Tixeront et al. (1951) working in Tunisia, varying between 0 and 53% for precipitation values between 300 and 780 mm.

Zebidi (1963), in Tunisia, studied a synclinal drain of limestone, extending over 19 km^2, and with a partial cover of maquis type vegetation and found an infiltration rate of 18%.

Bolelli (1951) found the following infiltration values for a Plio-Pleistocene limestone in Morocco:

Annual precipitation (mm):	500	550	600
Rate of infiltration (%):	14.0	16.5	19.5

5.2.2 Middle East

Burdon (1961) reports a rate of infiltration and re-infiltration of 41% in the Damascus basin (Syria) for an area of 5123 km² and with an average annual precipitation of 262 mm.

Issar et al. (1985) calculated the recharge to the Eocene limestone of the plateau of Avdat using figures for the spring and river discharge (900 000 m³/yr) and the potential recharge (average annual precipitation of 100 mm on a surface of 900 km²) and found a value of 1-2% of total precipitation.

Mero (1958) has found for the 200 km² drainage basin of middle Cretaceous limestone feeding the Na'aman spring in Israel that the average infiltration rate is 53% for an average annual precipitation of 600 mm calculated for the 1927-1957 period.

Shacori et al. (1965) made a series of observations in an experimental basin on the Carmel Mountain. The area is built mainly of limestones and dolostones of Cenomanian-Turonian age (Middle Cretaceous). Natural vegetation covers about 80% of the area. The average annual precipitation of 700 mm falls in winter. Observations on evaporation, runoff and water table fluctuations were incorporated in a balance equation. The results are given in Table 5.1.

Table 5.1 The hydrological balance of a basin in the Carmel Hills (mm) (Rosenzweig, 1972)

Season	Rainfall	Computed evapotrans.	Recharge	Surface runoff
1958/59	540	270	270	0.3
1959/60	650	380	267	3.0
1960/61	690	400	290	0.6
1961/62	760	300	460	0.5
1962/63	630	350	280	0.1
1963/64	760	420	340	1.3
Average	672	353	318	1.0

Rosenzweig (1972) carried out more detailed work in the same region, putting the main emphasis on measuring evapotranspiration by various methods. Taking into consideration that surface runoff is negligible, any amount not used by evapotranspiration is considered as groundwater recharge. The results show that in areas covered by a dense thicket of natural forest (mainly oak and pistachia) on limestones, evapotranspiration consumes the total amount of the precipitation, while in areas covered by annual grass (pasture land) the evapotranspiration is about 280 mm. The rest (420 mm or 60%) can be considered mainly as groundwater recharge.

Goldschmidt & Jacobs (1958), using empirical methods, made a study of the underground catchment of the Yarqon and the Nahal

Hattaninim (Israel), which are mainly composed of Cenomanian-Turonian limestone. From discharge values of the rivers they found that the average recharge is given by:

$$r = 0.86\ (p - 360) \qquad 5.2$$

where r : average recharge (mm/yr)
p : average precipitation above 360 mm/yr

The results of these works may give the erroneous impression that in arid regions where precipitation is less than 280 mm or 360 mm there is no recharge, yet this conclusion, as shown above, does not conform with empirical data. The reason for this is that in arid regions the extent of the vegetation cover on rocks is negligible and thus transpiration is marginal. In other words, it can be stated that the relation between precipitation and recharge above and below the threshold amount which enables vegetation is not linear.

5.2.3 Arabian Peninsula

Bahrain Wright et al. (1982) studied the hydrogeology of the islands of Bahrain on the Arabian Gulf. The islands forming Bahrain have an extremely arid climate. The rainfall is low and erratic, and comes mainly in the period November to April from a few heavy thundershowers and storms of short duration. The average rainfall is 72.5 mm, but is recorded to vary between 1.6-168.9 mm.

The main fresh water aquifers occur within the Alat/Neogene and Khobar carbonate formations. They are separated by a marly band. Both fissure and intergranular flow occur, but the former dominates, particularly in the top 10 m of the Khobar which is highly karstified. Therefore the Khobar has a higher hydraulic conductivity and forms the main supply aquifer.

Originally it was supposed that recharge would occur mainly at outcrops in the mainland of Saudi Arabia with additional leakage from the overlying Neogene and underlying deeper aquifers. The layer of fresh water floating on saline water in the interior basin of Bahrain is undoubtedly modern recharge water, as indicated by the hydrochemistry and the isotopic composition. It has been estimated that average annual recharge under present climatic conditions is likely to be in the order of 5-20 mm, which is consistent with the chloride content of the fresh water (100-140 mg/l).

Recharge of modern precipitation must also occur elsewhere in Bahrain, but conditions are generally less favourable than in the interior basin.

Qatar Lloyd, et al. (1987) summarized the extensive studies made of the hydrogeology of Qatar on the Persian Gulf. The climate is very arid with a mean annual rainfall of 75 mm in the north of the country and slightly less in the south. However, as in most arid zones the coefficient of variation of precipitation is very high and the mean is not a useful variable. Evaporation varies from 2.5 mm/day during winter months to 11.5 mm/day during the summer. Although the likelihood of direct recharge from precipitation through the soil profile is small, recharge of storm runoff often occurs. In Qatar the runoff recharge facility is provided by the numerous dissolution collapse depressions in the carbonates.

Studies of the hydrological balances of individual depression catchments resulted in relationships between storm intensity, runoff and recharge for individual storms:

$$Q = 188\ [(6.5 + p) + 8T]\ A \qquad 5.3$$

$$R = 124\ [(12 + p) - 3T]\ A \qquad 5.4$$

where Q : is runoff volume (m^3/km^2)
R : is recharge volume (m^3/km^2)
p : is rainfall (mm)
T : storm duration (min)
A : is catchment area (km^2)

The mean weighted runoff rate was 24% of rainfall. Recharge ranges from zero to 64% of rainfall in a storm event with a weighted mean of 15% as a percentage of storm rainfall. From these balance studies 10-12% of annual rainfall has been adopted for mean annual recharge in the resource assessment.

5.2.4 South Africa

Ghaap Plateau Smit (1978) studied the groundwater recharge to the dolomite of the Ghaap Plateau near Kuruman (South Africa). The average yearly rainfall on this area over the period 1940-1970 is 445 mm. This period can be divided into two parts: an above average precipitation (520 mm) occurred in 1949-1963, followed by below average rainfall (346 mm) in the period 1963-1970.

The surface area of the dolomite is about 1140 km^2. The rocks consists largely of dolomite of the Ghaap Plateau formation with lenses of chert and limestone. About 10% of the outcrop area is covered with sand and scree. The groundwater catchments normally differ from the topographic catchments. Numerous dykes divide the large compartments into about 50 smaller ones with the groundwater level falling stepwise towards the major Kuruman spring ("eye"), which has the largest flow. The water level in the dolomite varies between surface level and about 200 m depth.

The recharge to the dolomite aquifer was calculated by two different methods: (i) based on spring flow (variation) and (ii) based on climatic data using the Thornthwaite method.

In 1970 only three springs flowed and were gauged. The recharge can be calculated for the period 1963-1970 on the basis of the groundwater losses from the compartment (using the data given above), the rainfall and the surface area. The calculated average annual groundwater loss of 9 857×10^6 m^3 is supposed to be equal to the average annual recharge. When average annual rainfall for the period is 346 mm, the recharge is about 4% of this.

The Thornthwaite (1948) water balance method was only used for two sites with meteorological stations (Kuruman and Botitan). Using monthly average values for the precipitation and evaporatranspiration a zero value of recharge was determined, which shows that this method is not appropriate for these conditions.

Another approach was tried using the individual yearly rainfall pattern in relation to the average monthly evapotranspiration values. This showed that recharge takes place only during certain years of high rainfall. If the annual mean rainfall over the period 1940-1970 is used (445 mm), then the average yearly recharge at Kuruman is 3.39% of P and at Botitan 2.5% of precipitation. Both localities are in the same geographical area. An appropriate value for this area will be the mean, about 3%.

Large fluctuations in the individual annual recharge amounts were observed with values varying between 3-23% of P for Kuruman and 2-25% for Botitan. These values may be far more indicative than the average values.

Pretoria Region An area that forms part of a large karstic basin, extending from Pretoria to the western border of Transvaal (South Africa) was studied by Bredenkamp and Vogel (1970) and Bredenkamp et al. (1974). The results of these works will be given in more detail as they cover a variety of methods. The region is semi-arid. The mean annual precipitation is about 560 mm, which falls mostly in summer (October-April).

The main groundwater body is formed by the Dolomite Series, consisting of crystalline dolomitic limestone. The thickness of the dolomite, though irregular, increases with the dip of the substrata. It exceeds 1200 m in the central regions of the area under consideration. A number of diabase dykes subdivide the dolomite into separate groundwater bodies, and the study was focussed on the central compartment A, comprising about 224 km² and the adjacent compartment B to the west (Fig. 5.1). The topography of the area is remarkably uniform with a gentle slope to the west. There is virtually no drainage of surface water.

The soil cover of compartment A extends to a depth of up to 25 m. The soil texture is fairly uniform, varying from a sandy loam soil at the surface to a more clayey soil at depth. Outcrops of chert and dolomite constitute a relatively small percentage of the total area. The effect of fissures on recharge is thought to be small as they are partially filled by soil.

The surface of compartment B differs from A by having more extensive exposures of limestone and solid dolomite with occasional patches of soil and gravel.

The water is struck in fissures in the solid dolomite, ranging from small cracks to substantial solution channels (up to 10 m diameter). The cavities are often filled with chert or a honeycomb residue called "wad", which is formed during karstification and represents the insoluble ingredients of the dolomite.

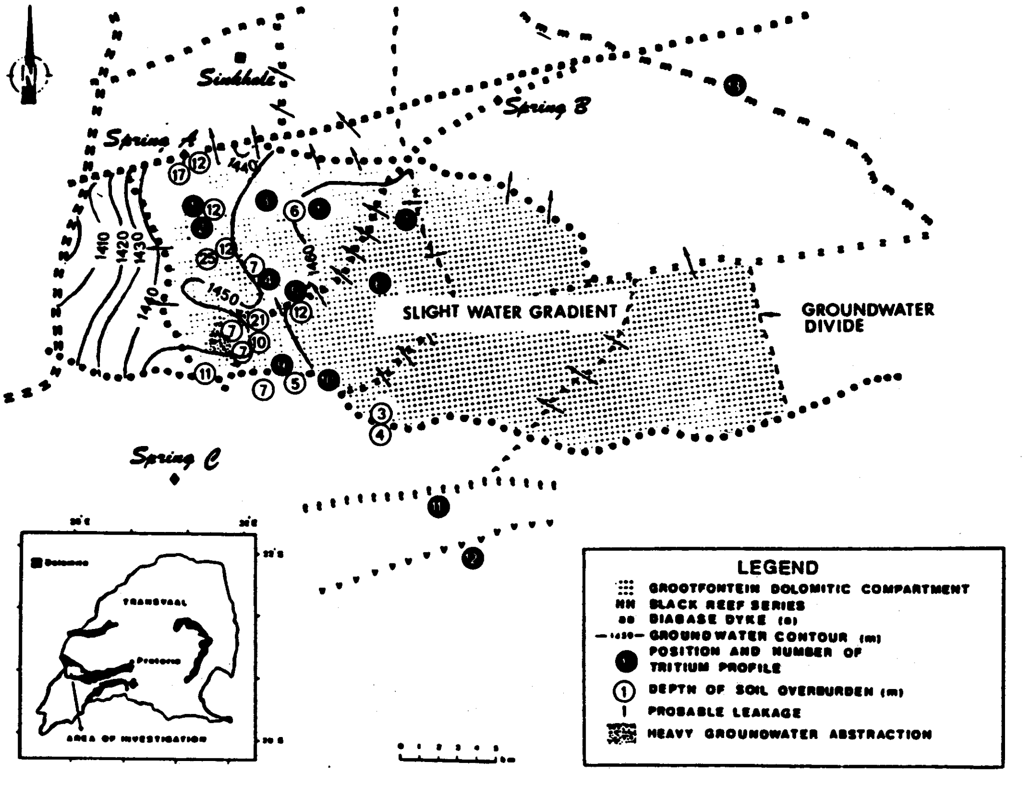


Fig. 5.1 Features of the Pretoria region karstic basin. Shows tritium sampling points, hydrogeological features, and probable leakages of groundwater through dykes (Bredekamp & Vogel, 1970)

The authors assumed that recharge occurs by a pistonlike movement of soil moisture through the unsaturated zone. As a tracer they used the bomb tritium peaks of 1958 and, especially, 1962/63. This method is especially useful for determining the recharge for an area with a homogeneous and relatively deep soil overburden. In compartment B it has been applied to deduce the recharge of a dolomitic aquifer even though the soil is not very uniform.

According to Bredenkamp and Vogel (1970), an indication of the occurrence of recharge is the clear linear increase in age of the water with depth, calculated from carbon-14 data.

The recharge is first calculated from the average yield of the major outlet, spring WT5. The average yield (458 m^3/h), divided by the surface area of the compartment (225 km^2) gives an average recharge of 17.7 mm/yr (3.2% of precipitation).

A full recharge inventory for the area over the period 1923-1968 can be constructed from the water level fluctuations in a sinkhole (Fig. 5.2). The righthand scale in this figure gives the actual value in rainfall equivalent, which is obtained by multiplying the lefthand scale with the specific yield (2%). The graph shows the frequency, magnitude and variability of the recharge.

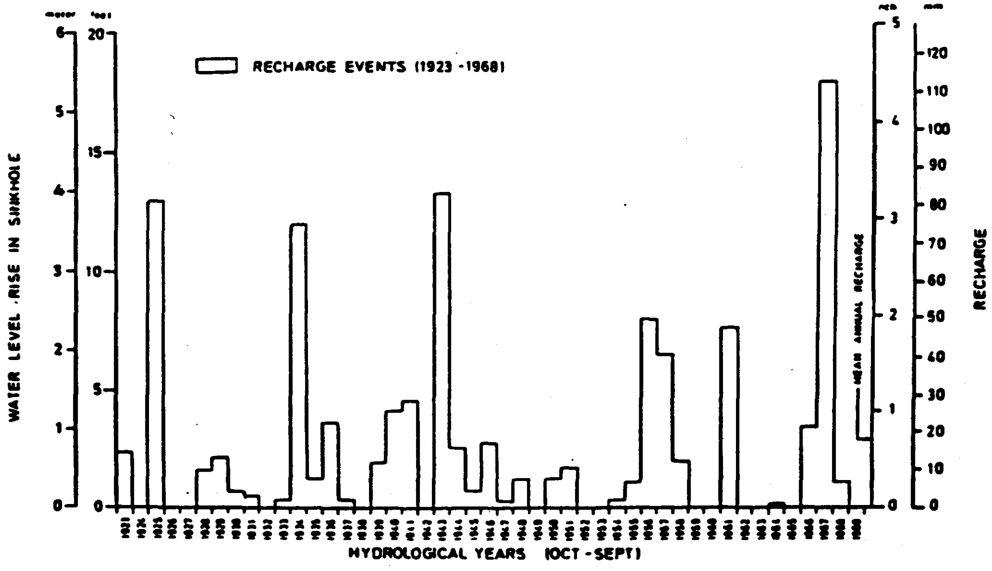


Fig. 5.2 Recharge events in Pretoria region. A reconstruction of events using the response of the water table in a sinkhole to rainfall. Right hand scale gives deduced water equivalent of the water table rise (Bredekamp & Vogel, 1970)

Using carbon-14 data, a rough estimate of the recharge can be made with the formula:

$$r = (n*D)/T \qquad 5.5$$

where n : porosity at the water table (2.73%)
D : average depth of the aquifer (47 m)
T : groundwater age determined with ^{14}C

From spring water sample WT23, located near WT5, it was found that T = 180 yr. Using this figure, r = 7.1 mm/yr (+2.5/-1.6 mm), which is about 0.98-1.71% of precipitation,p. At two other points the recharge was determined: (i) group WT22, 24 and 25: r = 3.3 mm/yr (0.6% of p) (ii) group WT27 and 28: r =

7.4 mm/yr (1.32% of p). The carbon-14 ages in this case have to be considered as orders of magnitude, as it is almost impossible to determine water ages to such an accuracy by this method.

These results are all clearly less than the value estimated from the average spring yield (17.7 mm/yr), but are not outside its range.

In their later publication Bredenkamp et al. (1974) used the same recession curve on a monthly basis to get a more accurate estimation of the recharge. For these data not only the time of occurrence but also the relative magnitude of the recharge events were reconstructed for the period 1923-1973 as shown in Fig. 5.3.

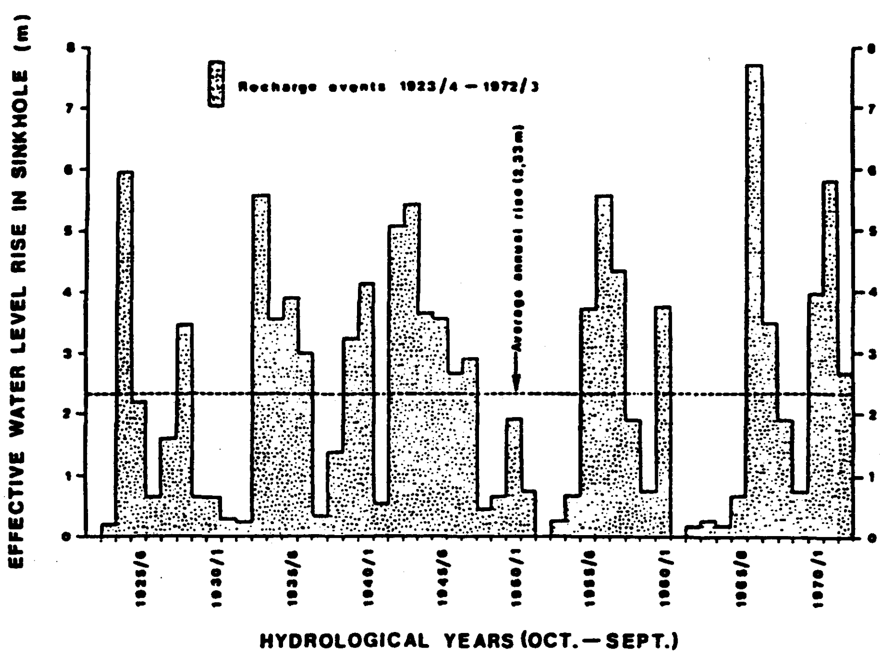


Fig. 5.3 Recharge estimates for Pretoria region, taking account of groundwater recession. A reconstruction of recharge events illustrated as an equivalent rise in water table (Bredenkamp et al, 1974)

In 1970, the authors reported that tritium was only found in the groundwater of compartment B where the soil cover is thin. Here the residence time in the unsaturated soil is short. In compartment A, bomb tritium water had still not reached the groundwater body.

The tritium results are plotted against depth below water table in Fig. 5.4. There is an exponential decrease in tritium content with depth (i.e. a linear increase in age of the water), assuming a steady state tritium input. The tritium content decreases by a factor of two (one half-life) in about 5.2 m, which gives for the recharge (0.027*5.2)/12.3 = 11.5 mm/yr (2.0% of p).

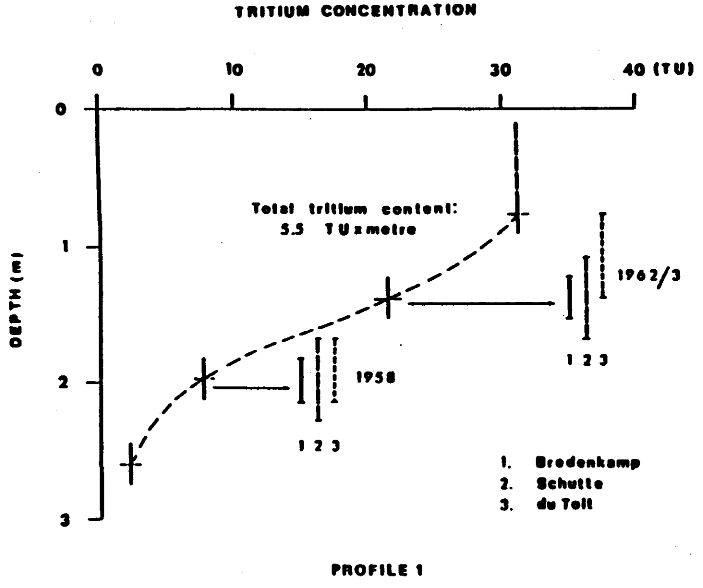


Fig. 5.4 Tritium profile from the Pretoria region. The probable variation in depth to which the 1958 and 1962/3 bomb tritium has penetrated is indicated for several independant interpretations (Bredekamp et al, 1974)

The tritium content of the spring can also be used to calculate recharge. The initial tritium content in the aquifer must, however, be known. By using an average age of 108 yr (+54/-29 yr), the recharge is (0.027*47)/108 = 11.8 mm/yr (+4.1/-3.8) (1.43-2.84% of p).

In a recent publication, Bredenkamp (1988) suggests a general equation for estimation of annual recharge in the Bo Molopo dolomitic region of Western Transvaal:

$$r_i = A\ (p_i - p_t) \qquad 5.6$$

where r_i : is annual recharge (mm/yr)
p_i : is the annual precipitation (mm/yr)
p_t : represents the threshold rainfall that is required to effect recharge, ranging from 310 to 360 mm
A : is a lumped catchment parameter, ranging from 0.28 to 0.35.

<u>Transvaal</u> Fleisher (1981) studied the hydrogeology of the dolomitic aquifers of the Malmani subgroup in the southwestern Transvaal (Republic of South Africa). The climate of the area of study is subhumid with typical summer rainfall and dry

winters. Mean annual rainfall decreases from 700 mm in the east to 590 mm in the west. A high potential evapotranspiration rate exists. Rainfall often occurs in the form of thunderstorms.

The Malmani subgroup is composed of predominantly dark grey dolomite with chert and quartzite beds and nodules. The dolomite mass is practically devoid of any primary porosity. The hydraulic conductivity is due to karstification. The aquifer consists of a jointed, fissured, fractured often karstic lower part and an upper weathered zone of decomposed rock material which includes a variety of fine to coarse fluvioglacial clastic deposits. Residual and transported chert breccias and conglomerates are the main water bearing components of the weathered zone. The aquifer follows the irregular surface of the underlying dolomitic bedrock, and is often overlain by a soil cover of varying thickness.

Water level fluctuations indicate that part of the natural recharge occurs immediately following precipitation. Spring discharge often shows a phase difference of four to six months. It is thus concluded that the natural recharge in the dolomitic aquifer involves, schematically, a two phase system: a relatively immediate phase, and a later delayed one. Percolation into the aquifer at various rates takes place throughout the year.

The rate of natural recharge as a percentage of annual rainfall was found to be between 13% and 27%. The lower percentages are typical for basins without any surface water courses. The higher percentages are often associated with favourable surface conditions combined with the occurrence of surface water channels and deeper water tables.

When the data from the above studies are put together (Table 5.2), there is an obvious trend of increasing percentage of recharge (R_{pc}) with increasing mean annual precipitation (p). In exponential regression form this becomes:

$$R_{pc} = A \exp(Bp) \tag{5.7a}$$

$$= 0.49 \exp(0.0045p) \tag{5.7b}$$

where A and B are the regression constants of the equation.

Table 5.2 Summary of data used to derive eqn 5.7b

Reference	Precipitation (mm)	Percentage recharge
Smit, 1978	346	2.5
Smit, 1978	445	3.0
Bredenkamp et al., 1970, 1974	560	3.2
Fleischer	645	13
Houston	937	30

The correlation coefficient for this equation is 0.975. Although this high value might be influenced by "spurious correlation" due to the very small amount of data and the large spread, an overall trend is clear. This correlation is remarkable as it has often been stated that recharge only occurs beyond a certain "zero" level of precipitation (see above on limestone aquifers).

5.2.5 Zambia

Houston (1982) used a recharge model based upon that developed by Penman for temperate climates. Modifications were inserted to account for time lag due to passage through the unsaturated zone and for the rapid inflow along fractures. Input to the model consists of rainfall and evaporation measurements and output or recharge is checked by comparison with water level hydrographs, which incidently leads to estimates of specific yield. Further verification of the model is obtained by comparison of these estimates of specific yield with those obtained from pumping tests.

The model was applied to a dolomitic aquifer in a semi-arid climate at Kabwe, Zambia. The aquifer is a Precambrian dolomitic limestone situated in a complex syncline and outcrops over 40 km^2. Flow within the aquifer is controlled by fissures. The average transmissivity is about 1000 m^2/d (max. 3500 m^2/d). The aquifer is phreatic, with very high values of specific yield (about 14%). The dolomite is overlain by permeable lateritic silts and sandy pockets usually ranging from 5 to 20 m thick. The original water table was found at a depth of about 10 m. The outcrop corresponds to a shallow topographic dome, surrounded by schists and gneisses.

The mean annual precipitation is 937 mm (over 70 years of records). Annual variation is high (640-1470 mm), with more high than low rainfall events. Rain falls mainly in the period December-March. The infiltration rates into the soil overlying the Dolomite Aquifer are frequently higher than 75 mm/h.

The evapotranspiration depends upon the vegetation cover and distinction is made between open forest (about 5% of outcrop area) and cleared forest areas, now used for agriculture.

After infiltration, loss occurs due to evapotranspiration. Evapotranspiration takes place at the potential rate unless the soil moisture deficit is smaller than the root constant, which must be estimated from the dominant vegetation type. The old (Penman) concept that no recharge occurs as long as soil moisture deficit exists is now found to be wrong: often filamentous (preferred paths) infiltration gives rise to recharge before the soil is saturated throughout.

It is supposed that all rainfall less than 75 mm/day is capable of infiltrating the soil. The rest goes to depression storage, of which 30% is assumed to infiltrate. Calculation over the period 1965-80 show that runoff is only 0.32% of precipitation.

Recharge was determined using three values for the root constant depending on the vegetation: (i) open forest (200 mm) (ii) short vegetation (75 mm) and (iii) bare soil (50 mm). Any water surplus is then taken to percolate downwards and recharge the groundwater storage.

The time step used was one month (due to lack of data), which is likely to lead to an underestimation of recharge, but it filters out the errors due to the time lag involved in percolation. The model was run over the period 1965-80, for which the average precipitation is 911 mm. As a result, annual recharge under bare soil varied from 26 to 771 mm (mean 281 mm) and under open forest from 0 to 534 mm (mean 80 mm). The results are shown on Fig. 5.5.

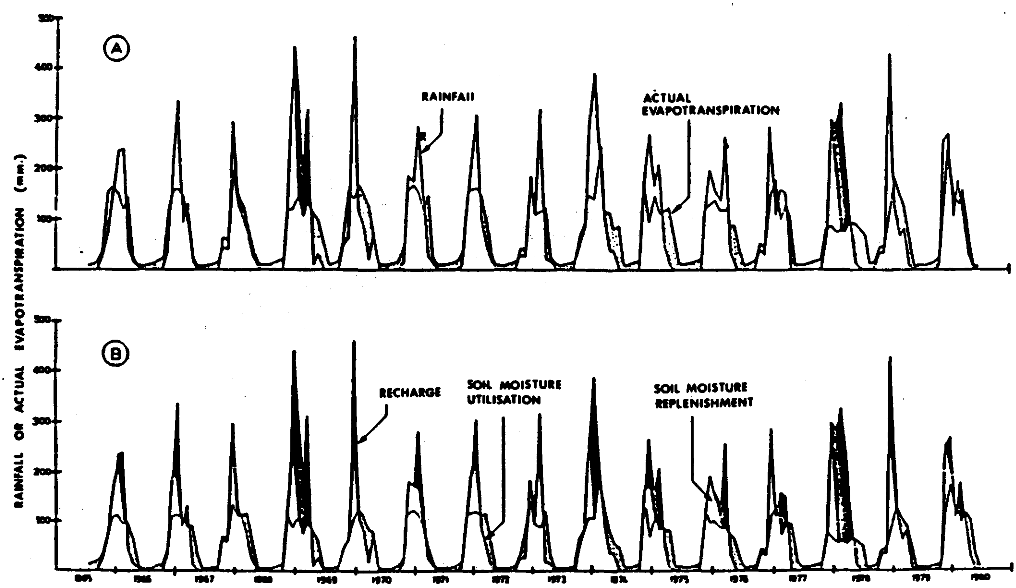


Fig. 5.5 Water balance for the Kabwe area, Zambia. (a) water balance under open forest, (b) balance under bare soil (Houston, 1982)

In his conclusion, Houston (1982) stated that the highest rainfall during the year in the months December-March coincides with low evapotranspiration and recharge can be expected to take place only during these months. The modification of the classic Penman method yields good results under semi-arid conditions provided attention is paid to the natural vegetation cover and consequent evapotranspiration values, and selection of an appropriate root constant.

5.2.6 South America

Salati et al. (1974) studied the Portiguar basin in northeastern Brazil. The Jandaira phreatic limestone is the second most important aquifer (first is the confined underlying Acu sandstone aquifer). Its mean thickness is 320 m. and its total surface about 18 000 km². This phreatic aquifer is fed mainly by rains falling over the region.

The authors claim that the differences in isotope composition (^{18}O, deuterium) between the aquifers show that upward leakage from the sandstone aquifer to the limestone aquifer is small (Fig. 5.6). This is also shown by the differences in ^{14}C ages. They quote the results of an hydrogeological balance for the area which estimated recharge to be 1.3% of annual average precipitation (Fig. 5.6).

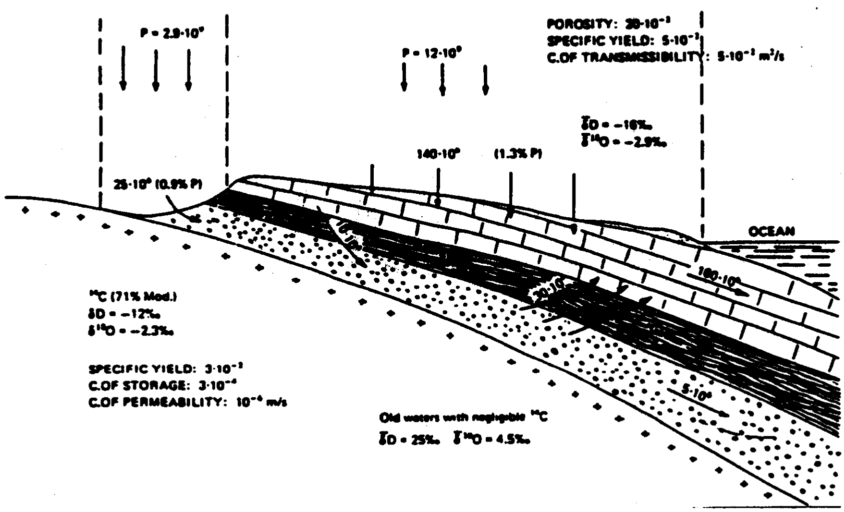


Fig. 5.6 Schematic water balance of the Potiguar basin, Brazil (Salati et al, 1974)

5.2.7 Australia

Allison & Hughes (1978) reported on a study of the recharge to an unconfined limestone aquifer of Oligocene-Miocene age in the Gambier plain in South Australia. Mean annual rainfall varies between 700 and 750 mm, most of which falls in winter. There is almost no surface runoff and drainage occurs through the soil to the underlying limestone aquifer. The area of the groundwater catchment is 1620 km². The soils of the study area range from the rather impermeable swampy podzols which can develop high soil moisture deficits, to the very permeable, skeletal soils of low water holding capacity which are usually only 50-200 mm thick and found developed directly on top of the highly permeable limestone. In an earlier publication, Allison & Hughes (1974) showed that local recharge is dependent on soil type. The area is divided into 10 different hydrologic units. The soils within each unit are considered to have similar hydrologic properties, and hence similar mean annual recharge (MAR). Irrigation is restricted to about 2% of the area. Approximately 15% of the total area is planted with Pinus Radiata. Earlier work has shown that direct recharge beneath these forests is much less than that beneath the surrounding grassland (e.g. Allison & Hughes, 1972). For the present study it was assumed that no recharge occurs beneath the forests. Sixteen samples were taken on improved pasture, which is the dominant land use of the area.

The recharge was determined using profiles of both tritium and chloride concentrations. It was assumed that the tritium concentration in the precipitation is constant over the area. For the use of chloride as a tracer it was assumed that the only sources of chloride are precipitation and fertilizer (potassium chloride). Exceptionally high chloride values of up to 300-400 mg/l could sometimes be found in the soil profile, while the water draining from the soil had a concentration of only 70 mg/l. This is thought to be due to salt sieving by clayey layers in the profile. Change in land use is clearly indicated by the differences in chloride level. When the recharge values found by the two methods are plotted together, good agreement results (Fig. 5.7). An important result is the apparent uniform recharge rate over large areas with similar soil type. The results for the different hydrologic units are summarized in Table 5.3. The total recharge over the area of study is 1.7x10^8 m^3/yr.

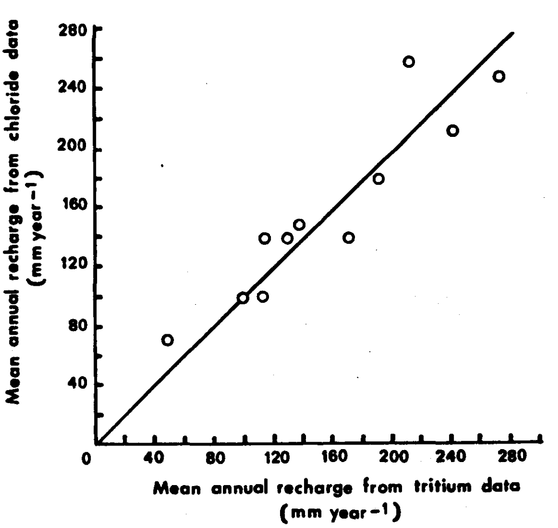


Fig. 5.7 Recharge estimates from chloride and tritium for the Gambier Plain, Australia (Allison & Hughes, 1978)

Allison & Hughes concluded that for the calculation with tritium data the sampling and analytical errors will become less important as recharge increases. However, since recharge is inversely related to chloride concentration, sampling and analytical errors will become more important at high recharge rates. Thus chloride should provide the best estimates of MAR at low values of MAR and tritium at high values of MAR.

Commander (1983) described the hydrogeology of the western Fortescue valley in northwest Australia. This valley, extending for 450 km, is filled at its western end with a sequence of dolomite, calcrete, gravel and pisolite which form an extensive aquifer system. The climate is arid. Mean annual precipitation is 350 mm, mainly from thunderstorms and tropical cyclones between December and May.

Table 5.3 Recharge through various hydrological units of the Gambier Plain, Australia (Allison & Hughes, 1978)

Hydrological unit	No. of sites	Area (km^2)	Mean annual recharge (mm) estimated using: Chloride	Tritium
Sand over heavy clay	1	157	70	50
Volcanic sands	1	21	100	100
Sand over sandy clay	2	52	140	100
Sand over thin sandy clay over limestone	2	49	105	120
Terra rossa over limestone	2	60	150	130
Thin sandy loam over limestone	2	281	140	155
Aeolianite	4	380	200	195
Skeletal soils	2	330	250	270

The Millstream dolomite is the most important and widespread aquifer. Recharge occurs mainly from floods, which is clearly demonstrated by the water table response. The 1975 flood caused a rise of 1.6 m which corresponded with an amount of 50×10^6 m^3. The variability of the runoff illustrates the difficulty of calculating a long-term recharge from this source.

Direct recharge to the dolomite aquifer can be estimated by use of chloride as a tracer. The rainfall contains about 1.1 mg/l. Near the groundwater divide, where only direct recharge occurs, the concentration of chloride is about 110 mg/l, which yields a recharge percentage of 1%. This represents less than 2×10^6 m^3 annually over the 540 km^3 of outcrop, and is therefore small compared with flood recharge.

6 CHALK PROVINCES

In general, it can be said that this is a complex porous, fractured, dissolved medium. The porosity is expressed in two coefficients. One is of the interstitial pores of the chalk itself which may change from that of clay to that of porous chalky limestone. The other porosity coefficient is that of the fine fracturing, especially along the planes of joints. A higher conductivity is found along the major fractures which are widely opened. Solution channels also develop along these fractures as well as along the main surface and subsurface drainage systems. It is usually assumed that the flow in these joints behaves according to Darcy's law.

Studies of the recharge into chalk have been restricted to a few regions, of which the London Basin Chalk is the best known example. Although the climate of this region is humid, it is supposed here that the characteristics of the London Chalk have universal application. This was confirmed by Issar et al. (1985) who studied the recharge into chalks in the arid zone and found characteristics that were quite similar to those of the London Chalk. Thus a brief description and main references for the London Basin will be given, even though this region does not fall into the category of arid regions.

6.1 The London Basin

The London Basin is a syncline in Cretaceous strata in southeast England (Fig. 6.1). The principal aquifer in the basin is the Chalk which outcrops on the northern and southern flanks of the structure and is overlain by Eocene strata in the centre (Fig. 6.2). The Chalk consists of a fine grained white limestone composed predominantly of fossil debris and micro-fossils (Anon., 1972). Measurements of porosity made upon samples of Chalk from southern England gave values of 25% to 45%. The porosity is both of primary and secondary origin. The intergranular pores are very small and water is held principally by capillary attraction. The specific yield is generally not greater than 5% and is probably no more than 1% or 2% at depth under the thick London Clay (Fig. 6.2).

The value of the Chalk as an aquifer is due to the development of secondary porosity in the form of intersecting horizontal and vertical fissure patterns. Generally the horizontal permeability exceeds the vertical permeability. Fissures are further developed by solution (Anon., 1972).

The Eocene comprises a group of sands and clays some 35 m thick, referred to as the Lower London Tertiaries, and an overlying clay, know as the London Clay, which is up to 150 m thick. A sand facies occurs at the base of the Lower London Tertiaries, which is commonly in hydraulic continuity with the Chalk (Fig. 6.2).

The Chalk attains a maximum thickness of about 250 m, but only the upper 50-60 m are fissured and thereby form the effective aquifer. The clays confine the groundwater under natural conditions, but nowadays the level is below the base of the confining layers (Downing et al., 1979).

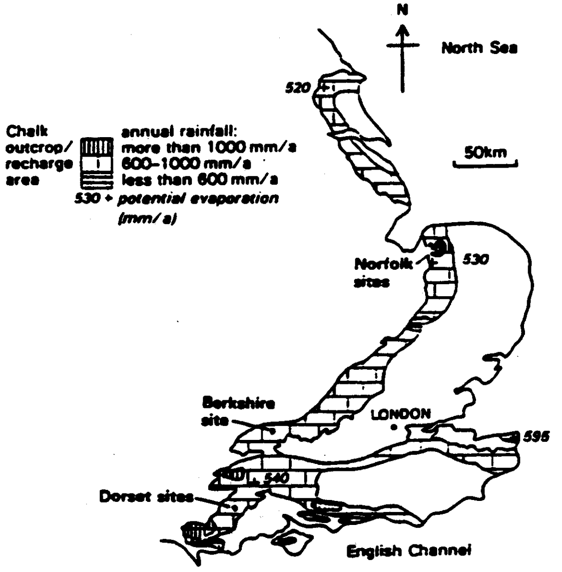


Fig. 6.1 Location and climate of the London Basin Chalk

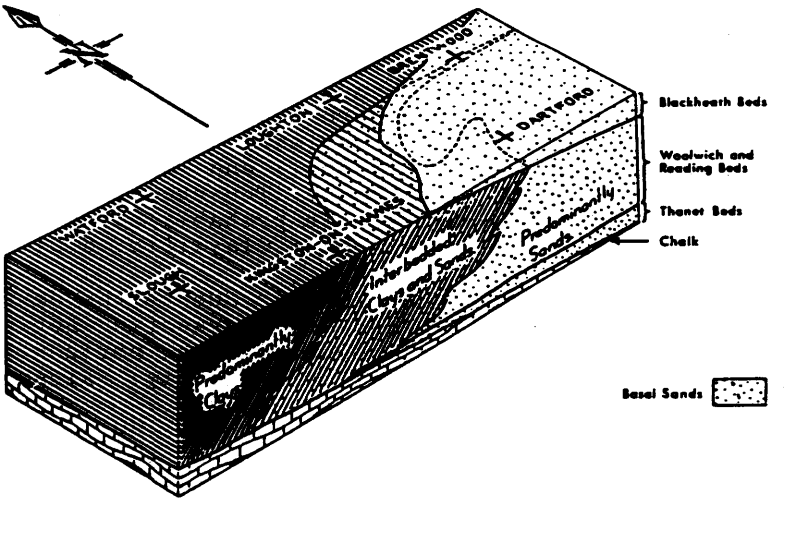


Fig. 6.2 Lithologies of the Lower London Tertiaries

The Chalks absorb rainfall evenly and the solution of Chalk by infiltrating rainfall does not tend to form the classic karst type topography which is more typical of hard compact limestones with low primary porosity. The solvent power of the infiltrating water due to CO_2 derived from the soil is largely depleted above the zone of saturation (Anon., 1972).

As the actual evapotranspiration over the Chalk outcrops is relatively constant, the areal variation of maximum available water for infiltration reflects areal variation in rainfall which is related to topography (Anon., 1972). Values for the maximum infiltration range from over 254 mm/yr in areas dominated by high ground to 152-178 mm/yr in the dip slope catchments. Where average rainfall is particularly low, infiltration may be only 25 mm/yr.

The Chalk of the London Basin is an extremely pervious rock and direct runoff from the outcrops is virtually nonexistant. Therefore, it can be assumed that where the Chalk outcrops the maximum amount of available rainfall infiltrates completely.

Where the Chalk is overlain by the Lower London Tertiaries, infiltration through this deposit, mainly through arenaceous sequences, recharges the Chalk. The assumption is made that infiltration amounts to 50% of the difference between rainfall and actual evaporation.

Oakes (1981) used a linear response function to investigate the delay of recharge to the water table from the infiltration. The infiltration was calculated from a soil moisture balance. In the method used it was assumed that infiltration at the surface in any month reaches the water table as a pulse distributed over a number of succeeding months giving rise to incremental water level rises. The linear response functions were coupled to an exponential recession of water levels to provide an auto-regressive, moving average model of water transfer through the unsaturated zone. The model was fitted to observed data to provide transfer functions for the infiltration process, and estimates of specific yield. The transfer functions were used to generate recharges by convolution with the surface infiltration. The observed lag was closely related to the thickness of the unsaturated zone and the speed of the infiltration pulse was estimated to be about 1 m/d.

6.2 Negev, Israel

A study by Issar et al. (1985) on the mechanism of recharge in an arid region built of chalk showed that permeable conditions may develop in the vicinity of ancient drainage systems, in agreement with the permeability distribution found by Lloyd et al. (1981). The permeability develops mainly along joints and fissures in the chalks. Tritium data have confirmed this statement.

The average tritium content of the various water resources in the limestone and chalk aquifers of the region was found to be a function of the depth of the water table, which is a measure of the storage capacity of the aquifer. A comparison was made between the tritium content of three springs discharging from the chalk aquifer. The largest springs with the most stable discharge (Ein Kudeirat and Ein Avdat) had values of 3-5 TU, while a smaller spring (Ein Aqev) with a more sporadic nature emerging from a little higher out of the same aquifer had an average tritium content of about 13.5 TU.

A similar pattern was found for the well of Nizzana, which is located in the chalks near the river bed of Wadi Nizzana. The depth of the water table is about 20 m and its tritium content is about 12 TU. At the well of Beerotaim, upstream on the same river bed, the depth of the water table is about 5 m and the tritium content is higher, namely about 22 TU.

Issar et al. noted that the fluctuations in water table and in tritium content of the water are often out of phase (Fig. 6.3). They explained this phenomenon by the following conceptual hydrogeological model.

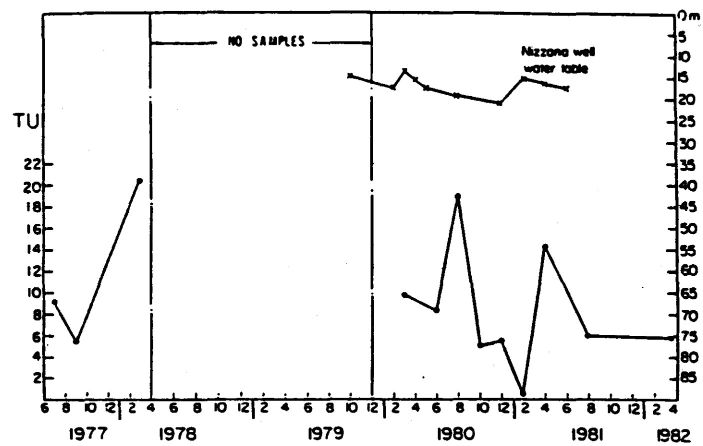


Fig. 6.3 Fluctuations of tritium and water levels in Negev chalks (Issar et al, 1984)

The groundwater in the chalks flows in solution channels developed along joints and fissures in the rocks (Fig. 6.4). These solution channels are mostly developed in the vicinity of rivers and stream channels. The larger and more ancient a system (two features which, in most cases, go together), the more developed are the solution phenomena. Thus, when a rainstorm occurs and the stream channels are flooded, the process of recharge to groundwater starts. This is first a pistonlike action on the more ancient water still contained in the solution channels, resulting in a rise in water table of the wells. Only in spring (and sometimes only in summer) does the newly recharged water arrive in the groundwater to affect its tritium content. This hydrogeological model explains some observations made in this region: (i) The gravel layers in river beds overlying chalk layers were found to be dry, while wells excavated in chalks, which are ordinarily impermeable, were found to contain water. It is interesting to note that wells of Byzantine time were excavated in chalks and not in gravel. (ii) In the same area and the same chalk layers, one exploration well had abundant water (about 20 m3/h), while another nearby well failed. The small quantity of water found in this well was more saline due to localized flow and the flushing rate.(iii) Salinity also increases with well depth because of low flushing rate. (iv) In an experimental spreading of flood water on the fields near kibbutz Revivim during the 1940s, a rise in the water table in a shallow well drilled in the chalks was observed simultaneously with a rise in salinity (from about 1000 ppm to 2000 ppm Cl).

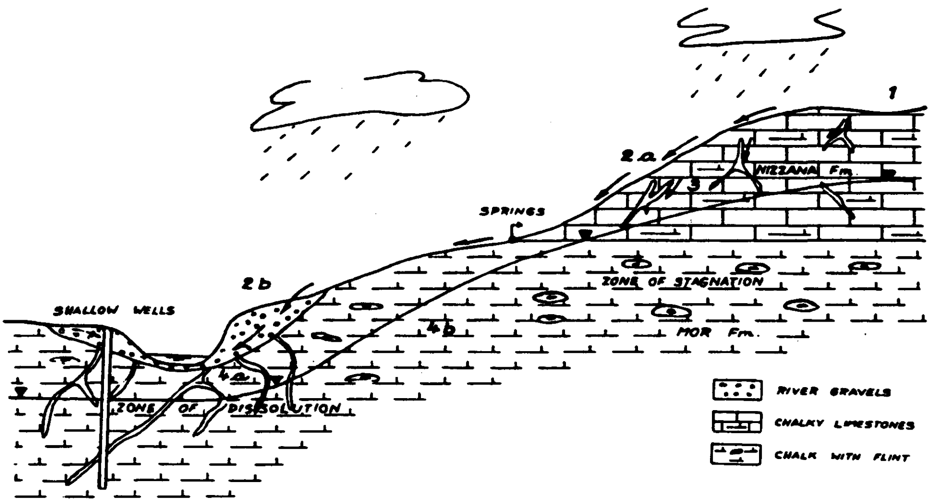


Fig. 6.4 Model of hydrochemical relationships in the Avedat group on the Negev Plateau

Notes:
(1) Precipitation: ionic ratios Ca>Na>Mg, HCO_3 >Cl>SO_4. Dominant factor distance from sea.
(2a) Runoff from small catchments: ionic ratios as above, enrichment factor 2-8. Dominant factor salt dissolution from soil and ion exchange.
(2b) Floods from regional catchments: ionic ratios as above, no significant enrichment compared to 2a.
(3) Groundwater in calcareous rocks: ionic ratios as above. Dominant factor salt dissolution from rocks.
(4a) Zone of dissolution, groundwater in chalks: ionic ratios Na>Ca>Mg, Cl>SO_4 >HCO_3. Dominant factor salt dissolution from rocks. High tritium values.
(4b) Zone of stagnation, groundwater in chalks: ionic ratios as above. Dominant factor salt dissolution. Low tritium values.

The overall recharge to precipitation ratio for this region built of limestones, chalky limestones, and chalks was found to be about 2.5% (Issar et al., 1985).

7 VOLCANIC PROVINCES

7.1 Plateau basalts

These cover extensive areas in Central India (The Deccan Trap), Central South America, and northeastern USA. The plateau basalts are characterized by the thickness of the basaltic lava and the lack of primary permeability. Thus their aquifer properties depend on secondary fracturing either due to release jointing or connected with faulting.

7.1.1 The Deccan Trap basalts

A summary of their hydrogeology is provided by Deolankar (1980). He concludes that the weathered part forms the best aquifer while the jointed basalts are the next best. The aquifers are of limited extent, suggesting the localized accumulation of groundwater.

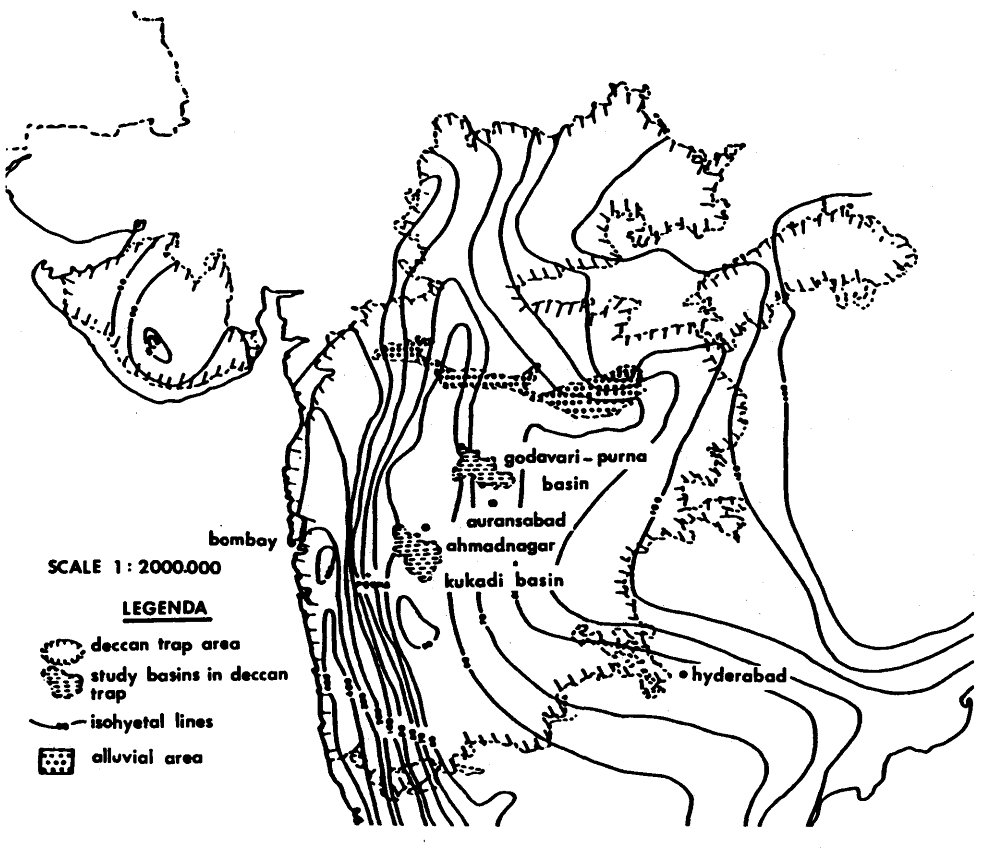


Fig. 7.1 Deccan Trap basalts, India. Shows isohyets and location of Kukadi and Godavari-Purna basins (Athavale et al, 1983)

The studies of recharge to the Deccan Trap basalts can be divided into two groups. One group comprises those which only use data from the groundwater in the basalt aquifer and the other group includes those which study the infiltration into the overlying (black cotton) soil.

Athavale et al. (1983) described the Deccan Trap basalts. This basalt plateau covers about 500 000 km² of western and central India (Fig. 7.1).

Physiographically, the area covered by the Deccan Trap formation can be divided into two broad units, separated by the continental divide of the Sahyadri mountains which run north-south nearly parallel to the coast.

The narrow coastal strip is characterized by heavy monsoon rainfall of 3000-4000 mm, while the larger plateau area, on the east of the continental divide, is situated in a rain shadow with average annual rainfall of 500-700 mm. Most of the rain falls in the monsoon period between June and September.

Two basins on the plateau were selected for the recharge study: the Kukadi basin comprising an area of 1153 km² and the Godavari-Purna basin of about 1091 km² (Fig. 7.1). Both basins are covered by basalt lava flows of Palaeocene age which are almost horizontal or dip gently. The lava flows are of two main types: compact massive or vesicular with or without amygdules. The thickness of individual flows varies from 5 to 50 m.

Both basins are covered with black cotton soils which are in situ. These soils were extensively studied by Hodnett & Bell (1981), and are discussed in Section 7.1.2. The thickness of the soil cover and weathered mantle varies between 1.5 and 9.5 m (average 5 m) for the Kukadi basin and between 1.5 and 39 m (average 14 m) for the Godavari-Purna basin.

Rainfall over the Kukadi basin for the months June 1980 to May 1981 was 612 mm and the annual open pan evaporation was estimated as 2226 mm. The average annual rainfall over 1980 for the Godavari-Purna basin was 652 mm and the open pan evaporation 1710 mm.

Tritiated water was injected at 19 sites in the Kukadi basin and at 24 sites in the Godavari-Purna basin during 1980, just before the start of the monsoon. Samples from the soil profiles at the sites of injection were collected in December 1980 from the Kukadi basin and in January 1981 from the Godavari-Purna basin.

The recharge at site was calculated by first determining the centre of gravity of the tritium versus depth profile. The mean displacement of the tracer was calculated as the distance between the injection point (60 cm) and the depth of the centre of gravity. The recharge was calculated from the moisture concentration, tracer displacement and wet bulk density.

From the profiles it was clear that the dominant displacement process is piston flow movement. Diffusion and dispersion of the tracer gives rise to attenuation and modification of the concentration profile.

For the Kukadi basin recharge varied between 135 to 0 mm; there were some discharges measured of up to 8 mm. The average of all values was 46 mm, which is 7.5% of the rainfall during 1980.

In the case of the Godavari-Purna basin the recharge values were found to vary from 208 to -28 mm with a mean value of 56 mm (8.6% of the rainfall of 1980). These values represent minimum recharge values as they do not include the contributions from stream and lake beds. The negative (discharge) values may indicate local (microscale) discharge zones or vertical upward moisture movement due to evapotranspiration or soil composition.

Using the assumption that slopes greater than 5% do not contribute to recharge, the total annual input is 31.9×10^6 m^3 for an effective area of 694 km^2 in the Kukadi basin and an annual input of 35.4×10^6 m^3 for 633 km^2 of effective area for the Godavari-Purna basin.

When compared with the recharge rate determined for the Neon basin with an area of 1085 km^2 using conventional hydrogeological and hydrometeorological methods, a figure of 32 to 36×10^6 m^3 was found for the year 1977. This figure compares well with the recharge figures found with the tritium method.

Another comparison is made between the recharge data from the tritium injection method and from water table level fluctuations in dug wells for 11 sites in the Godavari-Purna basin. The recharge to the phreatic aquifer can be calculated by multiplying the specific yield with the maximum water level change. A reasonable agreement was obtained for seven out of the eleven sites.

Limaye (1986), in a special report prepared for the present manual, stated that the Deccan Trap basalts can accept only a small percentage of the precipitation as recharge in areas of high rainfall. In areas having a rainfall in excess of 2000 mm, the recharge is as small as 40 mm. The rest of the rainfall contributes to runoff. A part of the recharge is later also lost to runoff as the bedrock outcrops due to the rugged nature of the topography. In the low rainfall areas, from the rainfall of about 250 to 300 mm the recharge can be as high as 100 mm. Due to a gentler topography, a thicker cover of weathered material and presence of permeable flow junctions, this recharges the groundwater. The residence time in the low rainfall region is up to 10 years for groundwater within the first 100 m of depth or so. In high rainfall areas the residence time is shorter.

For the semi-arid regions Limaye (1986) stated that a thick cover of black soil in some areas precludes any recharge from local rainfall.

The process of recharge is described using Figure 7.2. Hard basalt is exposed near the water divide and is overlain by a cover of soil and weathered rock in the central portion of the valley. Recharge in the rainy season takes place rapidly in the vicinity of the water divide where the soil cover is thin. The recharged water starts to move towards the centre of the basin in the fractured and weathered zone underlying the soil. In the rainy season considerable recharge also takes place from the runoff water flowing in the stream. A groundwater mound is then formed beneath the stream bed and it gradually spreads its base. This is called depression oriented recharge. The trough in the hard basalt below the stream bed is usually the zone of fracture concentration and recharge to permeable junctions between lava flows takes place through this zone. This water recharges deep permeable junctions between the basalt flows. The recharged water may travel longer distances than the water recharged along the water divide into the shallow fractures.

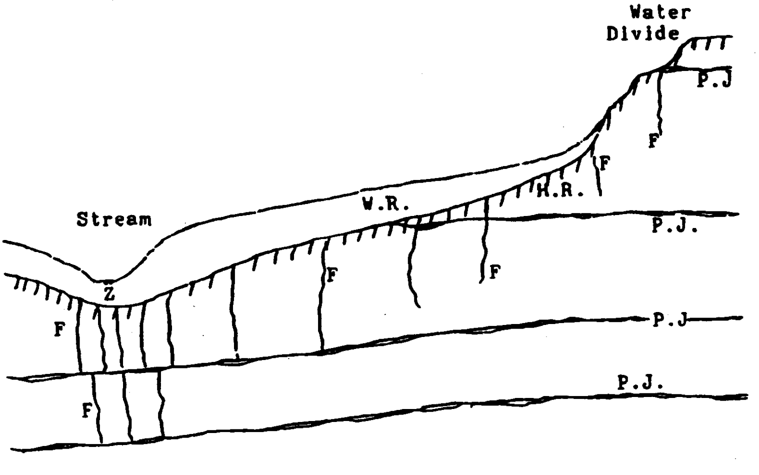


Fig. 7.2 Typical valley cross-section, Deccan Trap region (Limaye, 1986)

Notes: F - fissure, fracture or joint; H.R. - hard rock; P.J. - permeable junction between lava flows; W.R. - soil and weathered rock; Z - zone of fissure concentration.

Sometimes the trough in the hard rock and the zone of fracture concentration is located away from the present stream bed, especially in the case of rejuvinated streams. In this case recharge from stream beds is limited.

Limaye (1986) stated that the basalt of the Deccan Trap has special features. The primary porosity is low. The secondary porosity due to weathering is present in the near surface zone. Phreatic water occurs in the lateritic or alluvial cover and in weathered rock overlying the hard basalt. Below the surface of hard basalt semi-confined water occurs in the fissures, joints, fractures and permeable junctions between successive lava flows. As recharge has to pass through the soil before reaching the water table, the rate and amount of recharge is therefore largely governed by the hydraulic properties of the soil.

7.1.2 Black cotton soils

Hodnett & Bell (1981) studied the soil physical processes of groundwater recharge through Indian black cotton soils formed on the Deccan Trap basalts.

Black cotton soils are the major agricultural soils of a very large proportion of the 500 000 km^2 of the Deccan Trap basalt. They are dark coloured silty swelling clays which have developed on a light olive brown silty clay parent material ("yellow clay") derived from the weathering of basalts. Where the weathered basalt is within 2 m of the soil surface the black cotton soil usually overlies the weathered basalt directly. In the deeper soil areas the combined depth of the black cotton soil and yellow clay is generally between 2 and 10 m, but may even be deeper.

Black cotton soils show marked swelling and shrinking properties and an extensive pattern of cracks forms during the dry season. The cracks may be up to 75 mm wide at the surface and may reach depths of 6 m in the areas of perennial grass and shrub. Under cultivated areas 1.5-2 m is more typical. When very shallow, pieces of weathered basalt may occur and all basalt soils contain small concretions or "kankar".

Over most of the black cotton soil area, the water table is between 2 and 11 m below the ground level at the end of the dry season. In the monsoon it rises to within 0.2 m.

Hodnett & Bell (1981) concluded that in the area of study the black cotton soil is extremely uniform in its water holding properties and the depletion of its stored water in the dry season varies around 230 mm, the variation arising from crop type and localised winter rainfall. Depletion is insignificant below 2.5 m in the cropped areas and is mostly restricted to the top 1.5 m. In grass/shrub areas drying penetrates to twice this depth and depletions of around 500 mm are typical.

Deep drainage or recharge from the upper 2.5 m of the soil is only 30 mm (or less) during the dry season, from September to May. Most of this occurs during two weeks immediately after the monsoon. After that, the total drainage for the remainder of the dry season is 10 mm at the most, and probably much less. Exceptional winter rains can occur which completely refill the soil profile and thus increase this figure.

Evidence from the Dhàturi site, representing the deep soils, suggests that in these areas the structured black cotton soil clay forms an upper aquifer system which is virtually isolated from the more conductive weathered basalt aquifer by the "yellow clay". For these areas recharge to the deep aquifer is negligible during the dry season and certainly very much less than 1 mm/d during the monsoon. These soils only lose moisture by surface runoff, interflow and evapotranspiration.

For the shallow soil sites the black cotton soil forms one system with the underlying weathered basalt. This occurs when the soils are less than 1.5 m in the cultivated areas (deeper beneath grass/shrub). In most sites downward (recharge) fluxes are likely to continue throughout the monsoon.

A continuous pathway for significant input, as in the shallow soil areas, can form in the deep soil grass/shrub areas in those places where the weathered basalt is less than 4-5 m below ground level, provided there is capacity within the weathered basalt aquifer.

7.1.3 Idaho basalts

A thorough study of the recharge characteristics into basalts is described by Stephenson & Zuzel (1981). This paper reports on a recharge evaluation to determine the availability of groundwater on a grazing allotment on public land in southwest Idaho. As this study gives a very complete overview of the several factors influencing the recharge into basalt, it is summarized here.

The experimental recharge area is located on the semi-arid portion of the Reynolds Creek Experimental Watershed in southwest Idaho. The annual precipitation is 250 mm, with almost 85% occurring during winter and spring.

Geologically the area is primarily basalt, overlain by lake sediments at lower elevations. The basalt is underlain by granite at the upper elevation along the watershed perimeter. The soils are residual, developed primarily from weathered basalt, and range in depth from a few to about 60 cm. The loosely consolidated, basalt residuum grades from a clay-silt, sand-gravel matrix to altered fractured basalt at varying depths to a maximum thickness of about 3 m. Barren rubbly basalt outcrops of low relief comprise as much as 60% of the total surface of the study area. The vegetative canopy cover is generally less than 25%, but may be as much as 50% on the deeper soils.

Hydrologically, the site is located in the regional recharge area for the basalt aquifer system. Runoff events occur about once every two years on the average.

Out of a total of 14 observation wells, the data from 7 wells with the longest and most complete record were used to determine the recharge characteristics.

In their conclusions, Stephenson & Zuzel (1981) stated that:

(i) Groundwater recharge in these study areas occurs via several mechanisms under certain optimum conditions. Rainfall in excess of 20-30 mm over 24 h, or higher intensity cloudburst storms, are the major contributors to groundwater recharge. Infiltration and/or deep percolation to the water table follows.

(ii) Three separate mechanisms of recharge are identified in this semi-arid environment. Recharge from precipitation events can occur as water is transferred through low relief rubbly basalt outcrop areas, through the shallow soil residuum materials, and through the bedrock channels during runoff and channel flow.

(iii) Time to peak was found to be independent of season, depending only on soil depth.

(iv) The rate of groundwater recharge for the area, under the conditions given above, is estimated to be about 4.6×10^{-3} mm/min.

7.2 Vesuvian type basalt terrain

The character of aquifer systems in these terrains is determined by the alternations between pyroclastic (fragmental volcanic rock such as volcanic ash, tuff and bombs) and basalt flows; a function of the volcanic history of the region, namely, alternate episodes of lava flow and volcanic explosions. The time span between each active phase is important as it allows soil forming processes and the development of impermeable layers.

Due to the interlayering of lava, clastics and soil, many perched water tables can be found in the recharge zone. Chemical and isotopic composition of each depends on the lithology, mineralization and altitude. In volcanoes transversed by dykes, vertical compartmentation will further subdivide the terrain.

An important phenomenon in regions of large volcanoes is the recharge in areas of high altitude into the aquiferous perched lava flows surrounding the peaks.

7.2.1 Djibouti, Eastern Africa

Fontes et al. (1980) studied the groundwater systems of the Republic of Djibouti using environmental isotopes. The Republic of Djibouti is located in the arid zone of east Africa. Precipitation is randomly distributed on the coast and becomes monsoonal inland (summer rains). Temperatures are high and show little monthly variation.

The whole country is almost totally covered by lava and volcano-detrital deposits. The surface network at present consists of intermittent wadis. Groundwater systems are governed by geological and sedimentological features. Precipitation and/or floods can infiltrate through highly permeable fractured lavas and accumulate within sediments in the depressions, where porosity and permeability are extremely low due to the high concentrations of clay minerals.

Piezometric surfaces are very deep and range from 200 m below ground level in the North of the country to 30 m in the coastal region of Djibouti.

Isotopes were chosen for the unravelling of the recharge system because of the highly irregular pattern of rainfall data and the poor knowledge of evapotranspiration values.

Tritium measurements show values from background to 43±3 TU. Most samples thus contain a component of recent recharge. From the groundwater samples of the basaltic outcrops it was concluded that these aquifers are rapidly recharged through their fractures and cracks. Furthermore, the occurrence of recent water in the discharge indicates that the storage is low.

Some samples show tritium values which are at the lower limits of anticipated activity for recent groundwaters. These low values could be explained by a mechanism of recharge involving single episodes of tritium poor monsoon rains. This kind of recharge is compatible with local climatic and hydrogeological conditions, i.e. sporadic episodes of heavy rains and rapid infiltration into the fractured lava. Such a mechanism would also imply isolated circulation and low storage deposits.

Further observations on the recharge mechanism are: (i) The isotope values show that no significant evaporation occurred before or during infiltration, which is in agreement with the concept of rapid infiltration through fractured lavas. (ii) Isolated samples of rainwater fall below the local meteoric water line and are thus clearly evaporated during rainfall; this deviation shows that only heavy episodes of unevaporated rains can contribute to the recharge. (iii) Samples from the aquifer of Wadi Ambouli fall within a very narrow range close to the meteoric water line suggesting that recharge occurred from heavy unevaporated monsoon rains which generated heavy floods and recharged the aquifer.

8 CRYSTALLINE PLUTONIC PROVINCES

The recharge to these rocks is a function of two factors: the mode of chemical weathering of its surface and the rate of fracturing. As the chemical weathering produces soils rich in clay components, an advanced stage of such weathering may limit the recharge to outcrops which in the more humid regions may be rather limited. Lateritic processes may wash down the clay materials to form a perching layer between the upper soils and the subsurface.

The influence of fracturing on the permeability of crystalline rocks depends very much on their petrography and mineralogy, as well as the type of faulting. The more quartziferous the rocks the more brittle they are, and rather deep fractures may develop over wide areas. Tension faults give higher permeabilities.

In regions built mainly of crystalline acid rocks recharge can be rather high (up to 15% of average annual precipitation). The salinity and isotopic content will be that of the average rainstorms. Along deeper fractures higher salt contents may occur due to the dissolution of minerals.

8.1 Granitic terrains

Few studies exist on the recharge characteristics of granite terrain. Most of them form part of a broader study of volcanic and metamorphic rock hydrogeology. Only the parts dedicated to granite are described here.

Sukhija & Rao (1983) studied the recharge mechanism in the granites of the Vedavati river basin. The granites are coarse to medium grained, porphyritic and pink to gray in colour. The average annual precipitation is 616 mm and the evapotranspiration is about 71-77% of the annual precipitation. The gross annual recharge for the area was determined to be 13-21% of the annual rainfall.

The recharge values correlate well with age values of the water obtained from carbon-14 and tritium determinations. The investigation indicated that substantial recharge occurs along fractures and fissures. Major fractures have relatively younger waters as compared to areas characterized by small fractures and fissures.

Athavale (1985 and Chapter 18) described the injected tritium method for the determination of natural recharge and gave some data on the recharge in basins in India underlain by basalts, granites, gneisses and schists (Table 18.1). It is clear from these data that the recharge is normally about 8% of the average annual rainfall.

Muralidharan et al. (1988) report that a regional groundwater model, using a leaky aquifer concept, was prepared for the Vedavati basin in southwest India. The model showed a mean value of 42.5 mm for recharge. For the whole Karnataka state, in which most of the Vedavati basin drainage area is located, they estimate recharge to be about 34.4 mm or 5% of the average rainfall of 687.9 mm. This figure is in good agreement with the average recharge value of 39.2 mm obtained from tritium injection studies.

Allen & Davidson (1982) in a review of the groundwater resources in fractured rocks in Western Australia gave recharge estimations of several groundwater provinces consisting of granitic, volcanic and metamorphic rocks and gave values of 0.05-0.5%. The average rainfall for these areas is 300 mm or less. In general, potential evaporation exceeds precipitation by 3 to 10 times.

Thiery (1988 and Chapter 24) has investigated the mode of recharge into the fractured granites near Ouagadougou, Burkina Faso, Africa. A lumped parameter hydrological model was used. This model computes aquifer levels from rainfall and potential evapotranspiration data and is calibrated with observed wells. The monitoring well was drilled to 20 metres. It taps 5 m of granitic sand, 4 m of weathered granite and 5 m of fresh granites. The average annual precipitation from 1959-1985 was 825 mm but from 1978-1985 only 690 mm. Assuming a storage coefficient of 1%, the computed average annual recharge for the period 1978-1985 was between 23 and 45 mm/year, i.e., from 3.3 to 6.5% of the average annual rainfall.

Houston (1988 and Chapter 21) carried out an investigation on basement rocks in the province of Victoria in Zimbabwe, which included river baseflow analysis, hydrochemical analysis of groundwater and simulation modelling. All three methods produce consistent results suggesting that recharge amounts to 2-5% of annual rainfall (Table 21.3).

Issar & Gilad (1982) studied the hydrogeology of the arid crystalline province of the Southern Sinai (Egypt). The area has a rugged, mountainous topography with an elevation mostly above 1000 m. The mountains are incised by deep wadis. The crystalline basement is predominantly composed of Precambrian granitic, metamorphic and volcanic rocks intruded by many acid to basic dykes. The climate is typically arid with scant precipitation which occurs mainly in winter (November-March) when temperatures are relatively low. Precipitation is mainly controlled by topography, so the low coastal areas get about 15 mm/yr and the higher regions about 50 mm/yr. Maximum values reach 100 mm/yr.

A general evaluation was made of the order of magnitude of recharge to the crystalline fractured aquifers and the alluvial aquifers interconnected with them. For the catchment of the springs of Wadi Arbain, with an area of about 12 km^2, the average annual rainfall was 50 mm while the annual flow of springs and small wells amounted to about 90 000 m^3. From these data the recharge can be estimated to reach about 15% of the total precipitation. Part of the recharge is believed to find deeper and bigger fractures, and flows out of the region. A similar recharge value was found for Wadi Feiran. For the area draining to El Qaa at the coast from an inland catchment area of about 900 km² the average recharge was determined to be approximately 10-13% of the rainfall.

The establishment of the recharge and flow regime in this region was backed by a thorough study of the chemical and isotopic composition of the rain as well as the groundwater (Issar and Gat, 1981) (Fig. 8.1). From this study, it could be seen that there exist in the region several interconnected subsurface flow patterns. The shallowest is in the small fractures and is only a few metres deep. The water outcrops in small springs and seepages with a high tritium content (36 T.U.). Part of the water flows into deeper fractures which reach depths of tens of metres. Here the tritium content is low (7 T.U.). These fractures recharge and are recharged by the deep alluvial deposits of the oases (tritium content 3 T.U.). From the alluvium and deep fractures the water flows into the sedimentary basins bordering the granite province.

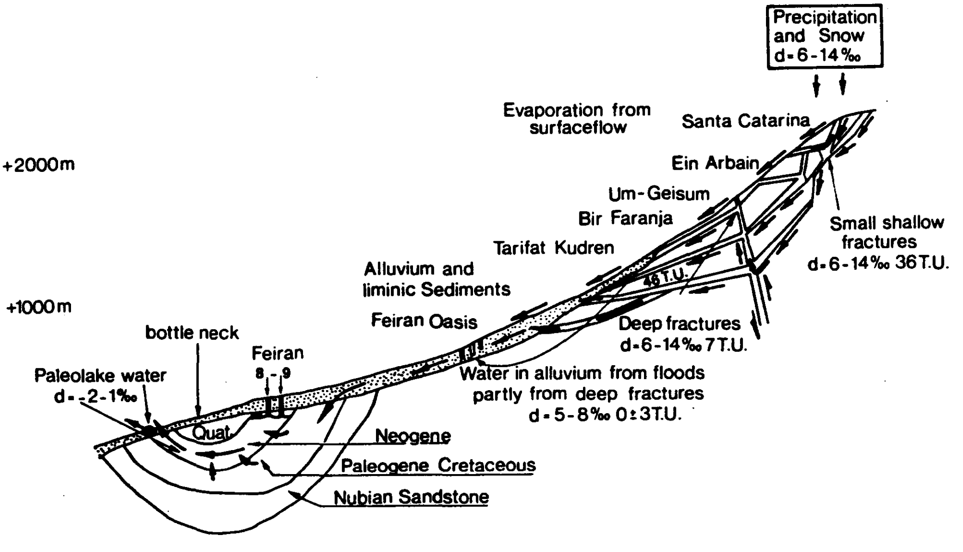


Fig. 8.1 Conceptual hydrogeological model of Wadi Feirian, Sinai (Issar & Gat, 1981)

9 A GENERAL PROCEDURE FOR RECHARGE ESTIMATION

9.1 The context of recharge estimation

Recharge is rarely estimated in isolation, but is usually just one aspect of a wider study such as groundwater resources, pollution transport, subsidence or wellfield design. It is easier, and more useful, to discuss a general procedure in a particular context, and this chapter uses the context of a regional groundwater resources study.

All groundwater resource studies are iterative because perfect data are not available and circumstances change in time (Fig. 1.8). This chapter discusses how to carry out one iteration, for example the preliminary evaluation, or the improved assessment that becomes possible once more data are available. In any such study there are five groups of work, the first two of which lead to a suitable conceptual hydrogeological model as discussed in Section 1.5 and Chapter 2:

(i) <u>Defining the groundwater system</u>, that is mapping the aquifers, non-aquifers, and interconnections and establishing the hydraulic nature of the boundaries. Mapping is essentially a geological exercise, but with a hydrogeological interpretation superimposed. Establishing boundary conditions is essential for modelling, but does not come easily to those with (hydro-)geological training. Fig. 9.1 defines the types of hydraulic boundary, and some discussion can be found in Engelen and Jones (1987).

(ii) <u>Defining the flow system</u>. This step identifies all the flows of water into, through, and out of the groundwater system. These are the recharges, discharges and internal flows. The mechanism controlling each flow must be identified; this is closely related to the identification of boundary conditions (Fig. 9.1) but also includes flow mechanisms within the aquifer and overlying materials - porous, fracture or dual porosity, homogeneous, stratified or multiaquifer, etc. Groundwater heads and gradients within and between units must be known to identify internal flows.

(iii) <u>Setting numerical values</u> for aquifer properties and flows and hence evaluating the water balance. Estimating permeability, storage and leakage factors is discussed in all hydrogeological textbooks, and of course this manual is concerned with estimating flows.

(iv) <u>Developing and proving a numerical model</u> of the groundwater flow system. This is the culmination of the steps above, and should proceed iteratively with them. A preliminary conceptual model and first estimates of numerical values are tested in a numerical model, which probably leads to revisions.

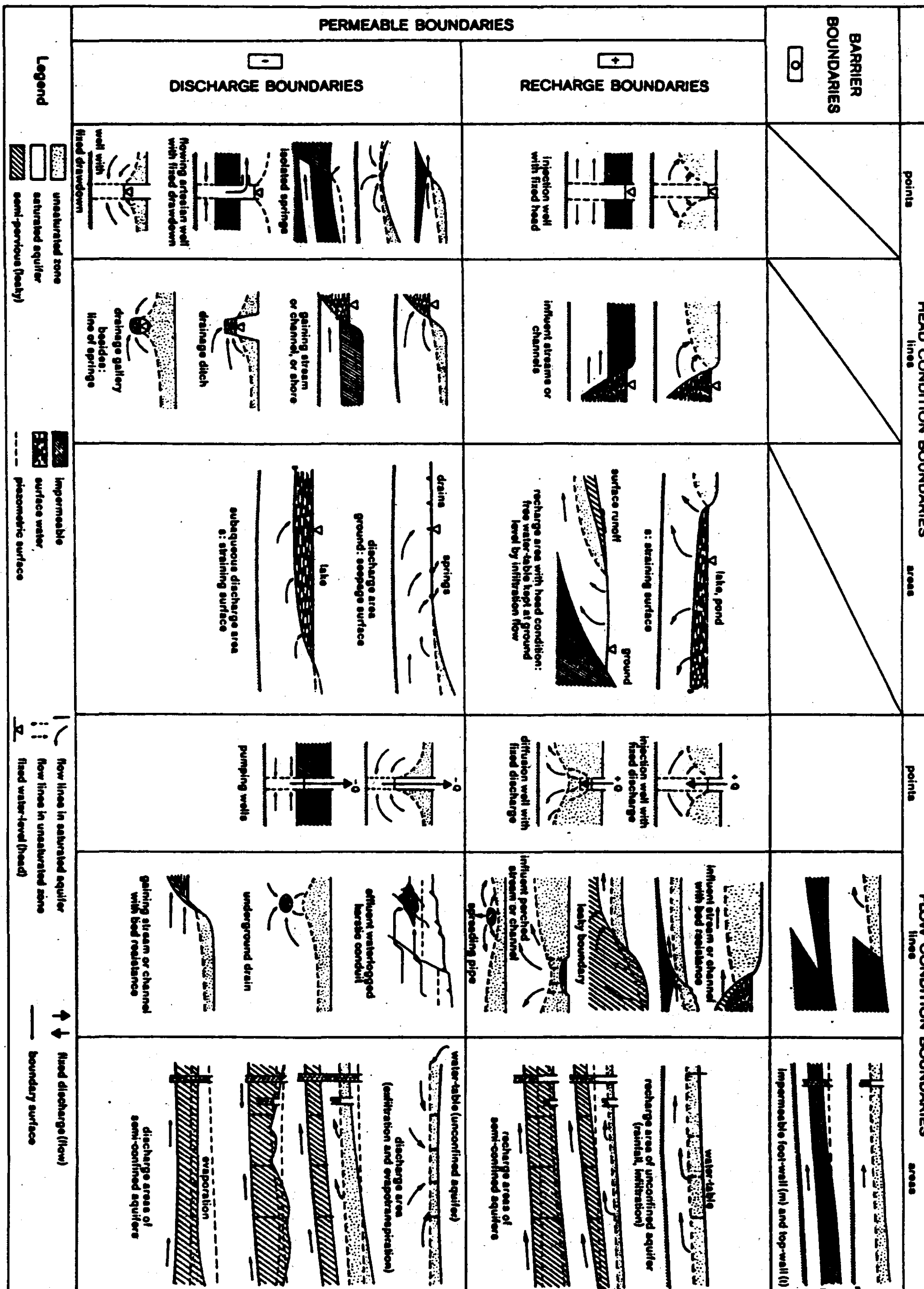


Fig. 9.1 Types of boundary conditions (Castany & Margat, 1977)

(v) Developing a groundwater resources plan is the last phase of work, and links to the socio-economic and financial parts of the study. It will use the numerical model to technically assess alternative plans.

9.2 A procedure

There are two strands to the procedure for estimating recharge which is set out in this section. The first is analysis of data from the region of interest. The second is comparison with other, hydrogeologically and climatically similar, regions. These may be anywhere in the world, but will fall in the same hydrogeological province as defined in Chapters 2-8. Transposition of data and understanding between comparable areas is often done instinctively by hydrogeologists who are used to working with insufficient data; it becomes easier as one gains more experience. Our proposal is that the process of comparison and transposition is formalised so that the maximum amount of information is extracted.

The procedure is outlined in Fig. 9.2, which should be read in conjunction with the more detailed description below.

9.2.1 Collect existing data

The first step in any study must be to collect all published and unpublished data and reports on the geology, hydrology, climate, pedology, etc. of the region. There are few regions of the world left where no data are available, although they may be buried in the filing cabinets of government agencies, foreign or local consultants, or universities. Many modern studies are too short to collect data on seasonal, interannual and longer time variations, and historical records are the only way to obtain such vital information.

9.2.2 Remote sensing and reconnaissance

If few data are available then rapid methods must be used to get the study underway. These may include use of Landsat and other satellite imagery, airborne remote sensing and aerial photography, and field or airborne reconnaissance studies. The objective is to define the groundwater and flow systems, as discussed in Section 9.1.

9.2.3 Conceptual hydrogeological model

Once all the readily available data have been gathered, a conceptual hydrogeological model must be defined. It should include climatic and hydrological aspects as well as lithology and groundwater. This will probably be revised (and refined) as more data are collected.

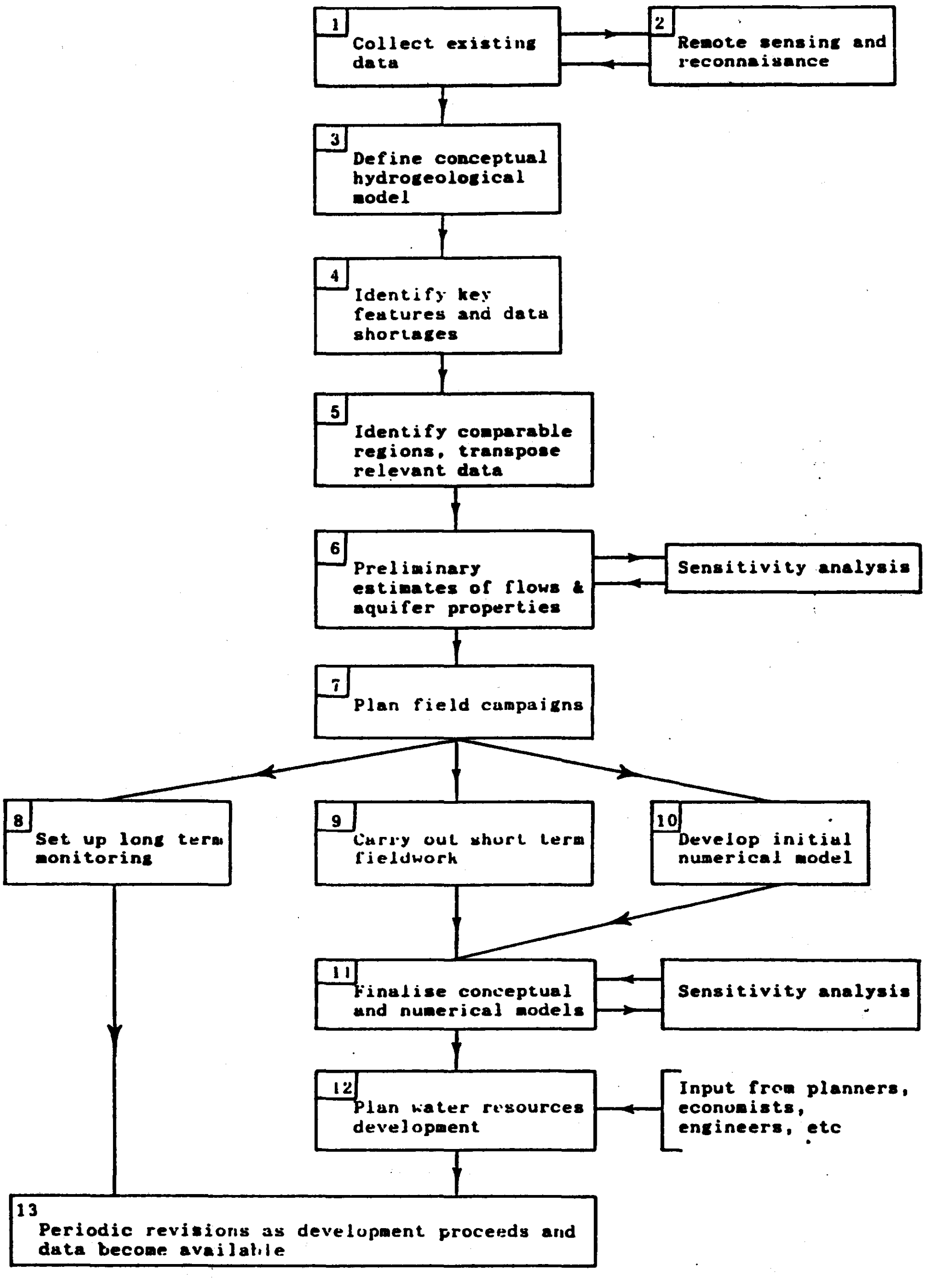


Fig. 9.2 Outline of general procedure for estimating recharge.

Note: This procedure is set in the context of a groundwater resources study. The figure should be viewed in conjunction with the text.

9.2.4 Key features and data shortages

Once the conceptual model has been defined, its key features should be identified. These will vary between regions, but will include:

- climate, particularly the type of precipitation events;
- the nature of the main hydrogeological units;
- flow mechanisms in aquifers and soil;
- internal and external hydraulic boundaries;
- expected piezometric patterns within and between aquifers;
- recharge sources, potential as well as actual, and their controlling mechanisms;
- discharges;
- degree of urbanisation;
- extent of irrigation, methods and water sources;
- level of exploitation of groundwater;
- expected hydrochemical patterns in the groundwater system;
- expected isotopic signals.

These key features can be used to identify comparable regions elsewhere, and can be checked in the field to confirm (or otherwise) the conceptual model. Data shortages, especially on the most important issues can now be identified to guide the planning of fieldwork.

9.2.5 Comparable regions

In a region of scarce data, it is often better to transpose data from a similar region elsewhere than to rely on guesswork. Hence we recommend that all available sources (journals, reports, personal knowledge, etc.) are searched to discover comparable regions for which more data are available. The key features of both regions must be compared to ensure that the similarity is close. Of particular concern are the degree of exploitation of groundwater and the use of the land surface. These factors can greatly influence flows, piezometry, hydrochemistry and isotopes, making it difficult to draw analogies.

9.2.6 Preliminary estimates of flows and properties

It should now be possible to make preliminary estimates of the essential numerical quantities: recharges, discharges, water balance, permeabilities, storage coefficients, etc. The objective at this stage is to see which items require fieldwork to improve estimates, and so a simple sensitivity analysis should be carried out to determine which are the most important, and which are the most poorly known.

9.2.7 Planning field campaigns

The scale of fieldwork that can be achieved within a project clearly depends on the time and money available: it is always possible to achieve something no matter how tight the budget!

More importantly, one must consider the longer term, as discussed in the following section.

9.2.8 Long-term monitoring

It is an irony of most project studies today that too little time is allowed to collect long-term data, yet the same region will probably be re-studied in a few years time, even if no water resources development has taken place. The moral is that a long-term monitoring programme should be started, even in short projects.

Such monitoring will be of flows and groundwater heads in areas which are expected to remain undisturbed as well as those where developments are likely. The latter, in showing responses to stresses on the groundwater system, will reveal much about the workings of the system.

9.2.9 Short term fieldwork

The fieldwork will depend on the data requirements of the study; see 9.2.4 and 9.2.6.

9.2.10 Initial numerical model

As well as being a predictive tool, a groundwater model synthesises all the available data and ensures consistency of interpretation. Modelling often reveals extra data shortages. Therefore modelling should be started as early as possible in any study.

9.2.11 Finalising models

To enable the final, planning, stages of the study, the conceptual and numerical models must be finalised. This will occur once all field data are collected and analysed. It should be accompanied by a sensitivity analysis to indicate the range of uncertainty in predictions by the model.

9.2.12 Water resource development plan

The culmination of the study will be a resource development plan. Planners and others will have inputs to this phase of the work, and the model will be used as a predictive tool. As discussed in Section 1.4, both the model and development plan will be revised as development proceeds and as long-term monitoring data become available.

9.3 Example - Recharge discharge ratios in the Nubian Sandstone aquifer of the Dead Sea region

During the early 1950s, the Dead Sea Works Co needed a reliable water supply, in the order of magnitude of a few million cubic metres per year to enable the production of potash. A few wells were drilled into the alluvium (Quaternary age) and limestone dolostone aquifers (Cenomanian age) which proved successful. A few years later the demand for water exceeded the safe yield of these aquifers. A search for an additional source led to drilling into the deeper sandstone aquifer of Lower Cretaceous age. The first exploration well struck an artesian flow. The piezometric head was found to be higher than that of the overlying limestone aquifer, while the salinity was lower (an average 1000 ppm Cl in the limestone, and 500 ppm Cl in the sandstone). It was thus clear that the two aquifers are separated in this particular region. In order to prepare a long term plan of pumpage from the sandstone aquifer, the recharge or inflow into this region had to be determined.

The collection of hydrological data from the region itself (mainly oil exploration wells) and from neighbouring regions (water wells and springs) enabled the formulation of two alternative conceptual models:

(A) Recharge from the limestone aquifer extending towards the northwest, underlying the southern Judean mountains, the annual average precipitation on which is 600 mm. The recharge was believed to be through the numerous faults crossing the region.

(B) Recharge from the Nubian Sandstone aquifer extending towards the south under the whole Negev and Sinai.

The key features of conceptual model A were those of a mountainous karstic limestone aquifer with high permeabilities in a semi-arid zone recharging a down-faulted sandstone aquifer of medium permeability.

The key features of conceptual model B were those of a sandstone aquifer containing fossil water extending over vast regions.

The comparable region for model A was the Judean limestone aquifer of the region of Jerusalem which was thoroughly investigated. The quantity derived by correlation with this aquifer gave an annual recharge of 10×10^6 m^3. The task was then to change the models developed for the other parts of Israel, to answer the special conditions of this particular region.

The comparable region to model B was the Continental Intercalaire aquifer of the Sahara and the Nubian Sandstone of the Western Desert of Egypt. Adopting the hydraulic coefficients known from these regions for calculating the annual subsurface inflow or recharge, the amount was found to be about 20×10^6 m^3/year. Another 70×10^9 m^3 was also found to be available as a one-time reserve which can be mined over a long period of time. The task in this case, was the development of a new model to investigate the response of an aquifer to extensive mining.

The investigation of the data from the two comparable regions showed that the most sensitive parameter was the isotopic composition of the water. While the water in the Judean limestone aquifer showed a contemporary isotopic composition, that of the Nubian Sandstone fossil aquifers was more depleted and had a lower deuterium excess coefficient (Issar and Gat, 1981).

The field data to decide between the alternative models was achieved by the collection of water samples from all wells and springs in the region, and the analysis of their isotopic composition. The results showed that model B is the one to be chosen.

A simulation numerical model has been applied. This involved:

(a) Definition of boundaries and boundary conditions (flow, water table);

(b) Division into cells by intersecting lines and columns;

(c) Definition of hydraulic coefficients T, S and thickness;

(d) Estimation of recharge-discharge for each cell (inflow-outflow);

(e) Preliminary water table map;

(f) Calibration between measured water tables at observation points and calculated water table, by changing parameters in a trial and error method.

This model enabled evaluation of a water balance, water table maps, hydrograph in various cells, and analysis of errors.

The preparation of the computer simulation model required a deeper analysis of the geological data. Vertical boundaries of the aquifer were drawn after preparation of the Lower Cretaceous structural map and a subsurface geological map of the pre-Lower Cretaceous unconformity. Horizontal boundaries of the aquifer were based on facies changes and structural barriers.

Transmissivity and storativity values were given to the aquifer according to hydrological tests and geological similarity in the areas where there are no drillings. The computer simulation model supported the main water flow assumptions with only slight alterations in the transmissivity values for areas with no existing field data, in particular the west coast of the Dead Sea into which a discharge of the Nubian Sandstone waters had been assumed.

However, the low water levels resulting from the computer modelling, when compared with the field data, pointed out a need for change in the transmissivity values in outflow areas of the Dead Sea.

The main conclusions were the following:

(i) Structural influences are important in affecting water flow within the Lower Cretaceous Nubian

Sandstone aquifer. The dominant flow direction is from the recharge area in the Sinai towards the Dead Sea and the Arava Rift Valley.

(ii) The northern aquifer boundary is set up by facies change from sandstone to a shaly aquiclude, while the southern boundary is influenced by the structural uplift that elevated the Lower Cretaceous Nubian Sandstone above the regional water level.

(iii) The major Lower Cretaceous Nubian Sandstone aquifer contains palaeowaters and is a distinctly different aquifer from the overlying Cenomanian. The sub-aquifer in the southern Rift Valley is hydrologically interconnected with the overlying strata, and therefore only one aquifer exists that transcends stratigraphic divisions.

(iv) No significant discharge of Nubian Sandstone water occurs along the west coast of the Dead Sea. Only small amounts of discharge are presently occurring at the southern margins of the Dead Sea.

On the basis of these conclusions, more wells were drilled towards the south while development towards the north was stopped.

A system of monitoring pumpage, water table levels, and salinity was put into operation. This also included data from the overlying limestone aquifer and neighbouring alluvium aquifer. A few wells (water and oil exploration) which penetrated the underlying Arad Formation (sandstone of Jurassic age) were also monitored.

In a more advanced stage of research, all data from the new wells as well as from the older wells were reprocessed by a more advanced numerical model. This model was based on the Integrated Finite Difference (IFD) method, the central concept of which was to divide the area under study into conveniently smaller subdomains, and to evaluate the mass balance in each subdomain. A regular network for the total flow region was constructed by applying the Thiessen method.

The numerical model was calibrated on the past behaviour of the system. Results from calibration of the aquifer flow model point to new aspects of the groundwater flow system:

(i) There is a connection between the Judea Group aquifer and the Kurnub aquifer in the fault region of the study area. Leakage is from the Kurnub Group aquifer into the Judea Group aquifer and the rate was found to be approximately $7x10^6$ m^3/year.

(ii) The Kurnub Group aquifer and the underlying Arad Group aquifer are connected. The leakage rate from the Arad Group aquifer into the Kurnub Group aquifer is approximately $6x10^6$ m^3/year.

(iii) A connection exists between the aquifers under study (Judean Group, Kurnub Group, and Arad Group) and the Dead Sea:

- The amount of water flowing into the Dead Sea from the Judea Group aquifer is approximately $15x10^6$ m^3/year.

- The amount of water flowing into the Dead Sea from the Kurnub Group aquifer is approximately $13x10^6$ m^3/year.

- The amount of water flowing into the Dead Sea from the Arad Group aquifer is approximately 10^6 m^3/year.

(iv) The main discharge of the aquifers to the Dead Sea is along its west coast.

(v) Leakage from the Judea Group and the Kurnub Group aquifers into the Arava gravel fill aquifer is approximately 20 x 10^6 m^3/year.

At this stage, a long-term exploitation plan, which divided the pumpage between the different aquifers, was prepared. This plan is rechecked every year and the model is recalibrated according to observations from the monitoring system.

Part III : TECHNIQUES

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10 GENERAL PRINCIPLES OF TECHNIQUES FOR ESTIMATING RECHARGE

10.1 Introduction

There can be several sources of recharge to a groundwater system. Each source must be considered separately to estimate recharge, except in some restricted circumstances discussed in Chapter 16. Therefore Chapters 11-15 of this part of the manual consider the main recharge sources individually:

11 - precipitation or direct recharge
12 - river recharge, including perennial, seasonal and ephemeral rivers
13 - interaquifer flows
14 - irrigation losses, both from canals and fields
15 - urban recharge

Each type of recharge can be quantified by several methods; there are similarities between methods for different recharges. The methods have been grouped into:

- direct measurement
- water balance methods
- Darcian approaches
- tracer techniques
- other, mainly empirical, methods

The remainder of this introduction discusses the requirements of a good method and a general approach to recharge estimation for resource studies. Sections 10.2-10.6 outline the common features of each group of methods; details are given in Chapters 11-16.

10.1.1 Requirements of a recharge estimation method

There are five essential ingredients of a "good" method:

(i) Water balance. A recharge estimate should explicitly account for the water that doesn't become recharge, so reducing the chance of over- or under-estimation.

(ii) Recharge processes. Only a few direct measurement methods do not rely on knowledge of the processes that convert source water to recharge and on flow mechanisms for that water. A good method will reveal if the conceptual model underlying the method is incorrect, for example if fissures sometimes carry bypassing flows.

(iii) Error of estimate. A good method will have low errors associated. It will not be sensitive to parameters which are hard to estimate accurately. For example, budgetting methods estimate recharge as a small difference between large numbers and so have inherently large errors. Unsaturated conductivity is difficult to estimate accurately. The problem of errors is discussed again in Section 10.7.

(iv) Ease of use. Methods which require expensive or unusual data, or specialized (non-hydrological) skills are not easy to use outside of research projects.

(v) Extrapolation. Recharge estimates are needed over long time spans for groundwater resource studies. Methods that can use readily available monitoring data, such as rainfall, to extrapolate estimates are more useful than those that require specialised observations (see also next section).

10.1.2 Time intervals

Each group of methods estimates recharge over a particular range of time intervals. Table 1.1 shows what time intervals are needed for groundwater resource studies. Shorter interval data can only be obtained by using other data to subdivide totals. Averages over longer periods should be obtained by summing values over shorter periods; because of non-linearities in recharge processes, and possible changes of

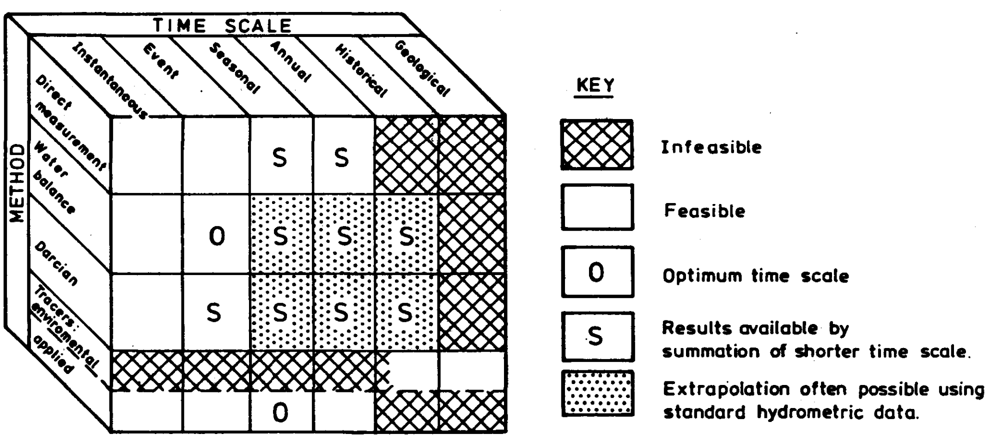


Fig. 10.1 Feasible time scales for estimating precipitation recharge, based on the methods available

Notes. *(1) The event time scale is marked as optimal for water balances because the conceptual model is most correct then, despite the higher errors of estimating such a small difference.*
(2) Applied tracers could be used on any time scale but are most accurate over a season.
(3) Optimal time scales are not shown for all methods.

flow direction, working directly with longer intervals may seriously mis-estimate recharge. An example is the use of soil moisture balancing models. Working with monthly data they usually indicate no recharge in arid and semi-arid areas, whereas using daily or event time steps, they will show that some recharge occurs. Fig. 10.1 shows the feasible and optimum time intervals for the various precipitation recharge methods.

10.1.3 General approach for estimating recharge

Recharge estimation must be seen as an iterative procedure. Initial estimates are revised and refined by comparing them to the results of other methods and to other data. Final estimates should result from calibration of a distributed groundwater flow model when groundwater responses in space and time are accurately matched.

The best initial estimates of historical recharge are obtained by considering (a) the recharge source, (b) the transmission route for recharge, and (c) the groundwater system. A general procedure is to ask the following questions:

(i) How much recharge can the aquifer accept? A full aquifer will reject further water which must then find another destination.

(ii) How much water can the unsaturated zone transmit? High potential recharge rates, for example from rivers or irrigation canals, may not be able to pass through low conductivity layers.

(iii) What other destinations are there for potential recharge, and how large are they?

(iv) How much potential recharge is there?

(v) What is the actual recharge? This step considers the balance and destinations of all water from the source, based upon (i)-(iv) above.

(vi) How do other estimates compare? Whenever possible, more than one method should be used.

10.2 Direct measurements

Recharge cannot be easily measured directly. The measurement must be at depth, when water has left the surface layers which can be under the influence of evaporation or other near surface processes. The intervening ground must be undisturbed so that the conversion of surface input to recharge is not affected by the measurement. Thus direct measurement requires construction and is expensive, particularly as it only provides a point measurement.

Only recharge from precipitation and canal losses are measured directly with any frequency, although the same techniques can be applied to recharge from irrigated fields. The construction and operation of lysimeters is discussed in Section 11.2, and the use of seepage meters is described in Section 14.2.

Direct measurement has the advantage of an implicit mass balance, that is all the input water can be accounted for. There are unlikely to be bypassing flows or overestimates of recharge. The time scale of measurements is that of the observations, from instantaneous up to seasonal or annual.

10.3 Water balances

This group of methods estimates recharge as the residual of all the other fluxes. The principle is that other fluxes can be measured or estimated more easily than recharge. Examples include:

(i) soil moisture budgets, in which rainfall and potential evapotranspiration are inputs to a soil moisture accounting procedure, with actual evapotranspiration and recharge as the outputs;

(ii) river channel water balances, when upstream and downstream flows are differenced to calculate recharge or - more accurately - transmission losses;

(iii) water table rise, when the volume stored beneath a rising water table is equated to recharge, after allowing for other inflows and outflows such as pumping wells and aquifer throughflow.

The advantages of water balance methods are that they use readily available data (rainfall, runoff, water levels), are rapid to apply, and they account for all water entering the system. Methods are available for all recharge sources; often they are the only feasible type of method.

The major disadvantage is that recharge is the residual, that is a small difference between large numbers. Errors can be high, with the errors in all the other fluxes accumulating in the recharge estimate. For example, high river flows can often only be estimated to ±25%. If recharge is 25% of flow, the error in estimating it is ±100%.

Other disadvantages include the difficulty of estimating other fluxes. For example, evapotranspiration cannot be measured easily, yet it is often the largest outward flux. Physical properties like specific yield are central to some water balance methods, but are not easily defined or measured (Section 16.2).

The natural time scale for water balance methods is the duration of a recharge event. Recharge processes are often non-linear, so that estimates based on longer time intervals should be summed over the individual events rather than calculated for the whole interval at once. Long records are available for much of the data used for balances (rainfall, runoff), so that long time series of recharge can often be calculated.

10.4 Darcian approaches

Groundwater flow is controlled by Darcy's law:

$$q = K\ i \tag{10.1a}$$

or

$$Q = K\ i\ A \tag{10.1b}$$

where
- q : seepage velocity (L/T)
- K : hydraulic conductivity (L/T)
- i : hydraulic gradient, $-dh/dx$
- Q : groundwater flow through a cross-section (L^3/T)
- A : cross-sectional area (L^2)
- h : hydraulic head (L)

When combined with an equation of mass conservation, this gives an equation of flow, for example, in the x-y plane below the water table:

$$\frac{\delta}{\delta x}\left(T\frac{\delta h}{\delta x}\right) + \frac{\delta}{\delta y}\left(T\frac{\delta h}{\delta y}\right) = S\frac{\delta h}{\delta t} + q(x,y,t) \tag{10.2a}$$

or for vertical flow in the unsaturated zone above the water table:

$$\frac{\delta}{\delta z}\left(K(\theta)\frac{\delta h}{\delta z}\right) = \frac{\delta \theta}{\delta t} + q(z,t) \tag{10.2b}$$

where
- T : transmissivity (L^2/T)
- x,y,z : coordinates (L)
- q : outflow per unit area (L/T)
- S : storage coefficient
- t : time (T)
- θ : moisture content
- $K(\theta)$: unsaturated hydraulic conductivity (L/T)

Knowledge of some of hydraulic heads, pressures, moisture contents, hydraulic conductivity and other properties and boundary fluxes should, in principle, enable recharge to be estimated. There are two broad approaches; field measurement and numerical modelling.

Field based techniques usually assume steady conditions, when only measurements of head (or pressure) and hydraulic conductivity are needed to apply eqn 10.1 or its unsaturated equivalent. The methods work well for saturated flow, provided conductivities can be measured at the right scale - laboratory values often bear little relation to large scale field values. Unsaturated flow is much more difficult to calculate from field measurements because of the sensitivity of conductivity values to moisture content (Fig. 11.8).

Numerical modelling methods take transient flows and storage changes into account, and can include spatial variability of physical properties, of which hydraulic conductivity is the most important. However data requirements and computing load are high.

The principal advantage of these methods is that they attempt to represent the flow of water - the actual physical processes that we are interested in. This is often countered by the need to make simplifying assumptions in order to reduce the computational effort. For example, numerical models of the soil zone usually assume a single porosity medium with no spatial variation in properties. In practice many soils have dual porosity, with preferred pathways during high satuation, that is at times of recharge.

The correct time scale for such models depends on the rate of fluctuation of heads, varying from seconds for rainfall into soil, to seasonal or longer for seepage between aquifers.

10.5 Tracer techniques

Both environmental and applied tracers are widely used for recharge estimation in arid and semi-arid areas. Environmental tracers can be used in both signature and throughput methods. Signature methods are when parcels of water are dated or labelled. Throughput methods involve a mass balance of tracer. Applied tracers are normally only used in signature methods; lateral dispersion makes it difficult to achieve a mass balance. Tracers are widely used for precipitation, irrigation and total recharge studies, but not for quantifying river or urban recharge.

Because tracers do not measure water flow directly a number of problems can arise, leading to over- or under-estimates of recharge. These problems include secondary (unknown) tracer inputs, mixing and dual flow mechanisms, and only arise if the sources, sinks and pathways of tracer are not fully understood. Part of the recharge going through preferred pathways such as root channels or fissures may invalidate a tracer method. Ironically tracers are often used to detect such dual flow mechanisms.

Environmental tracers sum recharge over long times in arid and semi-arid areas. This can be tens of years when profiling in the unsaturated zone, to thousands of years for aquifer wide tracer balances. Applied tracers are used where recharge is higher and work over one or several seasons.

10.6 Other methods

A miscellany of other methods is in use for recharge estimation. These are mainly empirical, in which recharge is correlated with other variables (precipitation, elevation, canal flow). The relationship is then used (a) to extend the recharge record in time, or (b) is transposed to other catchments of similar characteristics.

In many practical studies such transpositions are the only way to proceed, especially for preliminary estimates of recharge.

It is clear that catchments should be closely matched in characteristics, and that the empirical relationship is only as good as the recharge estimates on which it was based. A difficult problem is the effect of groundwater levels on recharge; high or perched groundwater levels will reduce recharge, sometimes by several orders of magnitude. Depth to groundwater is the most difficult factor to correlate between catchments, and so empirical methods work best when the water table is deep and no perching occurs. Changing groundwater conditions (once resources are exploited) may change recharge, but empirical methods cannot estimate these changes as they contain no model of recharge processes.

Non-empirical methods used in particular circumstances include hydraulic models of river flow (Section 12.6.3) and inverse modelling methods for total recharge (Section 16.4).

10.7 Accuracy of recharge estimates

10.7.1 Types of error

Because recharge is not easy to measure directly, estimates of it are prone to large errors. Four types of error can be identified and are discussed below:

- incorrect conceptual model (defined in Section 1.2)
- neglecting spatial and temporal variability
- measurement error
- calculation error

Conceptual model. This is the most serious and most common type of error. It arises when the recharge process is not fully understood, or simplifying assumptions are made. For example, it may be assumed in a particular case that excess irrigation water applied in parks becomes recharge, while in truth a low conductivity layer causes perching and horizontal flow to a surface drain. As another example, a monthly time step might be used for a soil moisture budgeting model in a semi-arid area resulting in zero recharge being estimated, whereas occasional short wet spells overcome soil moisture deficits to cause some recharge.

It is impossible for the authors of this guidebook to identify and catalogue all the ways recharge can occur. All we can do is continually emphasise the need to use all the available evidence, understand the system you are working with, and to point out the most important simplifications in the various methods.

Temporal and spatial variability. Most recharge processes are non-linear in relation to time. For example a low intensity of rainfall may cause no recharge because of a high rate of evapotranspiration, whereas the same amount in a shorter time period may be sufficient to saturate the soil and cause recharge. Thus errors will arise by ignoring temporal variations, for example by using monthly, annual or long-term average data. Recharge is similarly non-linear in respect of spatial variations of inputs and physical properties of soils and aquifers.

Measurement error is governed by the equipment used to make measurements. It can be estimated by experiment, can be handled mathematically (Section 10.7.2) and is the kind of error most usually considered. Handbooks on hydrological methods often quote the proportional error associated with standard measurements.

Calculation errors can be avoided by care and checking, especially of units. A particularly difficult type can occur with numerical computer models unless they are vigorously tested under a wide range of conditions.

10.7.2 Theory of errors

Measurement errors are the only type which can be examined mathematically. They can be expressed as a proportional error (p) or as an absolute error (a); for example, river flow might be measured to within a proportional error range of ±10%, or within an absolute range of ±10 m^3/s.

Errors should be given as the standard error of estimate. That is if the variable X is being measured it can be expressed as $(X_t + e_x)$, where X_t is the true value and e_x is the error on an individual measurement. X will be normally distributed with mean X_t and variance a_x^2. The shorthand notation for this is

$$X \sim N\,(X_t, a_x^2) \qquad 10.3$$

The errors are also normally distributed

$$e_x \sim N\,(0, a_x^2) \qquad 10.4$$

as are the proportional errors

$$e_x/X_t \sim N\,(0, p_x^2) \qquad 10.5a$$

$$p_x = a_x/X_t \qquad 10.5b$$

It is a property of the normal distribution that 68% of values will lie within the range $(X_t - a_x,\ X_t + a_x)$.

With these definitions of error it is possible to consider how much error arises from the use of different kinds of equation. Consider an additive equations such as a simplified river channel water balance

$$R = U - D - E \qquad 10.6$$

where R : recharge rate $[L^3/T]$
U : flow in river at upstream section $[L^3/T]$
D : flow in river at downstream section $[L^3/T]$
E : evaporation from river surface $[L^3/T]$

Replacing the estimated values by true values and errors gives

$$(R_t + e_R) = (U_t + e_U) - (D_t + e_D) - (E_t + e_E) \qquad 10.7$$

where subscript t indicates a true value and e is the error in each variable. Assuming that U, D and E are independent, normally distributed variables, the error in R is given by

$$a_R^2 = a_U^2 + a_D^2 + a_E^2 \qquad 10.8a$$

or, for proportional errors

$$p_R^2 = p_P^2\ (U/R)^2 + p_D^2\ (D/R)^2 + p_E^2\ (E/R)^2 \qquad 10.8b$$

Two conclusions can be drawn from eqn 10.8b for additive equations. First, the relative size of the variables must be known in order to handle proportional errors. Second if, as is usually the case, R is small compared to the other variables, the proportional error in R will be large; in fact it can easily exceed 100%.

Consider now a multiplicative equation, for example Darcy's Law (eqn 10.1b). In this case, the error in Q is given by

$$p_Q^2 = p_K^2 + p_i^2 + p_A^2 + p_K^2 p_i^2 + p_K^2 p_A^2 + p_i^2 p_A^2 + p_K^2 p_i^2 p_A^2 \qquad 10.9a$$

The last four terms are one or two orders of magnitude smaller than the first three, that is

$$p_Q^2 \approx p_K^2 + p_i^2 + p_A^2 \qquad 10.9b$$

and for absolute errors

$$a_Q^2 \approx i^2 A^2 a_K^2 + K^2 A^2 a_i^2 + K^2 i^2 a_A^2 \qquad 10.9c$$

Thus the sizes of the variables are needed to use absolute errors with multiplicative equations. Eqn 10.9b is the most useful, and should be compared with eqn 10.8b. It shows that multiplicative equations lead to much smaller errors than additive ones.

10.7.3 Minimising uncertainty in recharge estimates

Error analysis, or sensitivity analysis by differentiation, will show which variables in an equation lead to the highest errors, and effort can be concentrated on obtaining the most accurate estimates for these. However this approach will not help if the conceptual model is wrong. It is strongly recommended that more than one method of estimation using other data is used to provide an independent check.

Calibration of a groundwater flow model against observed groundwater responses should always be used as the final step in recharge estimation (see Section 10.1.3). A groundwater flow model may allow prediction of isotope and other tracer concentrations in the groundwater system and its outflows. These predictions can be compared to observations, providing another check on the overall conceptual model and recharge estimates.

11 PRECIPITATION RECHARGE

11.1 Introduction

11.1.1 Direct and indirect recharge

A simplified view of the precipitation-recharge process is shown in Fig. 1.2. In essence, some of the precipitation returns to the atmosphere by various evaporation processes, some runs off laterally, and the remainder becomes direct recharge. This chapter is concerned with estimating this direct recharge, that is recharge below the point of impact of the precipitation. Including moisture storage above the water table:

$$\text{recharge} = \text{precipitation} - \text{runoff} - \text{actual evapotranspiration} \pm \text{storage change} \qquad (11.1)$$

An immediate conceptual difficulty is, "How far can water move sideways and then infiltrate before counting as indirect recharge?". Any such movement implies variability in spatial properties which can invalidate many of the estimation methods for direct recharge. On the other hand, small and numerous movements cannot be counted or treated individually when building a regional groundwater model. Examples include:

(i) weathered, bare, hardrock or limestone terrain where recharge is into distinct fissures,

(ii) surface depressions, with sizes and spacings from centimetres upwards, where local runoff gathers and infiltrates,

(iii) bare rock outcrops in permeable terrain (eg sand or alluvium), where runoff infiltrates at the edge of the outcrop,

(iv) the many minor, ephemeral drainage channels with permeable beds which may only contribute to main channel flows on rare occasions; at other times all flow infiltrates into the bed.

These examples illustrate that a pragmatic division into three is needed, (i) direct recharge at the point of impact (ii) indirect recharge along main watercourses, and (iii) an intermediate category of recharge where some horizontal, surface flow occurs but not involving a mapped water course. Most of this chapter deals with (i) direct recharge; Chapter 12 deals with (ii) indirect recharge. We have called the intermediate category *localised recharge* (see Section 1.1.1). It is the least well researched of the three. The last section of this chapter (11.8) discusses how some information relevant to modelling studies might be obtained.

11.1.2 Methods for estimating direct recharge

Methods for estimating direct recharge can be classified as follows:

(i) Direct measurement over areas up to 100 m^2.

(ii) Totally empirical methods, which usually simplify eqn 11.1 to:

$$\text{recharge} = f(\text{precipitation}) \quad 11.2$$

where the function may be linear or non-linear and may involve easily measured variables like altitude or catchment area.

(iii) Water budget methods, (i) at a point or (ii) at a catchment scale. The former are usually soil moisture budgetting methods.

(iv) Darcian approaches, that is making use of the equation of flow in the zone above the water table. These may be based on field measurements of moisture and head, or on numerical models.

(v) Environmental or applied tracers which track the movement of parcels of water in the unsaturated zone.

These methods are discussed in some detail in Sections 11.2-11.6, and a brief summary of their appicability, ease of use, costs, etc is given in Table 11.1. Sections 11.7 and 11.8 deal with variability of recharge across catchments and localised recharge respectively.

Potential and actual recharge. A useful conceptual distinction is between actual recharge which is the water that reaches the water table, and potential recharge which is available, but which may go to another destination. The clearest example is with a high water table, when potential recharge (excess of precipitation over evapotranspiration) cannot enter groundwater and becomes runoff. If the water table subsequently fell, more actual recharge would occur from the same potential recharge.

Possible destinations for potential recharge are runoff, evapo(trans)piration, storage in the soil or unsaturated zones, perched water tables or minor aquifers, and discharge through local groundwater systems instead of the regional system. Whether these apply to any particular groundwater system will depend upon the boundaries adopted.

Actual recharge is needed when simulating historical conditions of a groundwater system. However, when modelling possible future conditions, for example when new boreholes have lowered the water table, it may be more appropriate to use potential recharge. Many methods of estimating recharge estimate potential, rather than actual.

Table 11.1 Comparison of methods of estimating precipitation recharge

This table is not a substitute for the longer discussion of the methods given in the main text of this chapter.

Method	Lysimeters	Soil moisture budgetting	Numerical modelling of unsaturated zone	Field-data based Darcian methods	Tracer profiling methods
Section	11.2	11.4	11.5.1-11.5.4	11.5.5-11.5.6	11.6
Applicability	All situations where construction is feasible	Humid, temperate or irrigated areas, with soils. See Eqn 11.7	Isotropic soils without cracks, swelling clays, root channels, etc	Limited to special cases See text	All tracers - unsaturated zone only, water table well below root zone, single flow mechanism. Environmental - hydraulic steady state. Applied - enough recharge for measurable movement
Accuracy	Good	Moderate to poor in semi-arid, worse as more arid. Should be calibrated (Section 11.4.8)	Theoretically good. Must be calibrated	Poor to moderate in unsaturated zone, moderate to good below water table	Moderate to good, if applicable
Data requirements	All collected on site	Daily meteorological data, more frequent in more arid areas. Knowledge of vegetation and cropping patterns	Daily meteorological data. Soil properties (K-p-θ relationships). Calibration data (eg soil moistures or pressures, gw levels, lysimeters)	Regular soil moisture and pressure measurements. K-p-θ relationships	Input history for environmental tracers. Tracer and soil moisture profiles by sampling
Ease of use	Construction - good engineering contractor and careful supervision needed. Operation - competant technician, daily monitoring	Simple arithmetic	Generally considered as research tools. Preparation of computer code, soil property data, and use of models can be difficult and time consuming	Fairly simple	Chloride profiling is fairly simple. Isotopes require special expertise and equipment
Type of estimate	Point value. Only for period of observation, ie usually short term	Point value, or estimate for homogenous zone	Instantaneous point value	Point values over period of data collection	Point value, averaged over period of profile
Costs	High. $5-20 000 for installation, plus monitoring	Low if data already available	High. $5000+ for soil property measurements, high manpower costs.	High. $2-10 000 for installation, plus monitoring	Moderate. $250-2500 per chloride profile, plus $250 per tritium analysis
Time required	Long to settle in and for data collection (>1 yr), longer in arid and semi-arid areas (>5 yr?)	Rapid if data available, otherwise as long as period of observation (several years)	1-6 manmonths if data and modelling skills available, 6-24 manmonths if not	As for lysimeters	Moderate. Applied tracer methods require at least one wet season, possibly several years

11.2 Lysimeters - direct measurement

The only practicable method of measuring recharge flux is with a lysimeter. This is a block of soil instrumented so that flows through it can be measured. The block is isolated from the surrounding soil but is representative because it has the same vegetation and climatic exposure. In order to minimise edge effects and average out local variations in soil and vegetation lysimeters need to be large (up to 10 metres in each of the three dimensions), and so are expensive to construct. Examples of their construction and use for recharge measurement are given by Kitching & Bridge (1974), Kitching et al. (1977), Kitching & Day (1979), Kitching et al. (1980) and Kitching & Shearer (1982).

11.2.1 Design requirements

A lysimeter design for estimating recharge should fulfill the following requirements:

(i) Contain undisturbed soil. Vertical flows through repacked soil will not be representative.

(ii) Be large enough to minimize edge effects and average small scale heterogeneities. A minimum plan area would be 1 m^2, the ideal would be 100 m^2.

(iii) Be large and deep enough to enclose complete root systems. This sometimes makes naturally vegetated lysimeters impractical, especially in arid and semi-arid areas where roots can be 50 m deep.

(iv) Be surrounded by similar vegetation to avoid oasis effects.

(v) Have the same hydraulic condition at the base as found at the same depth in the surrounding soil. This is discussed more fully below.

(vi) Be watertight, except for the drainage to be measured.

11.2.2 Hydraulic conditions at the base

Vertical flow in the lysimeter will be partly controlled by conditions at the base and so these must be the same as in the surrounding soil. If the aquifer has a shallow water table, a water table should be maintained in the lysimeter at the same depth. If the aquifer's water table is deep, suction must be applied at the base of the lysimeter. There are three base conditions in use, illustrated in Fig. 11.1:

(i) A water table lysimeter in which water is pumped in or out to maintain the same level inside and out. Fig. 11.1a illustrates such a lysimeter on a thin outcrop of aquifer, where an impermeable base is formed by an underlying clay unit.

(ii) Suction drained lysimeters have a base above the water table with suction drainage installed just above the base (Fig. 11.1b). In order to keep water potentials and hydraulic gradients in the lysimeter the same as that in the surrounding soil, the suction applied should either be based on the depth to water table or the suction measured in the surrounding soil. The total drainage is measured and is equal to recharge.

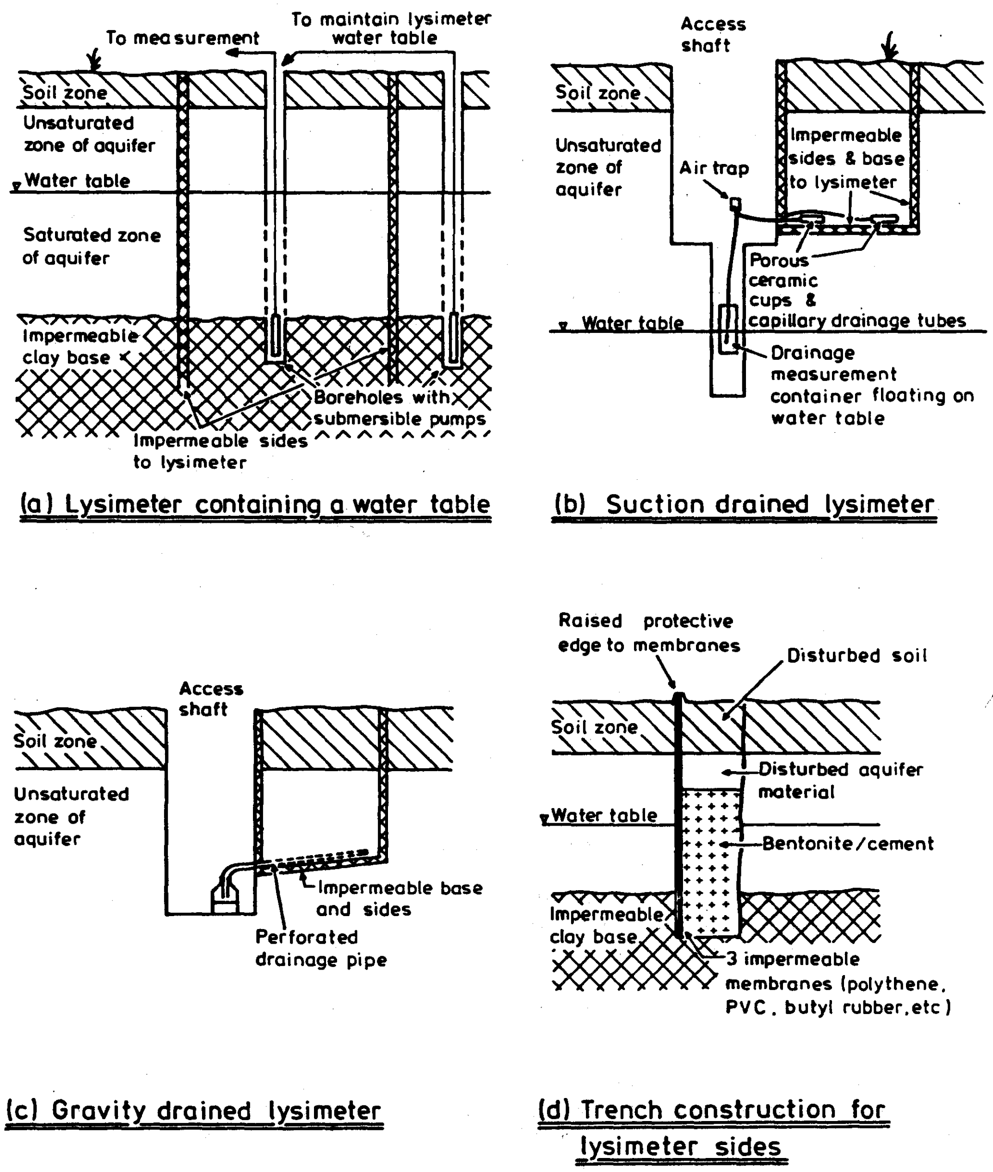


Fig. 11.1 Design and construction of lysimeters

(iii) A gravity drained lysimeter (Fig. 11.1c). These are only of use in coarse grained, permeable soils where no significant capillary rise will occur and where conditions at depth have little influence on vertical flows.

11.2.3 Lysimeter construction

The sides of large monoliths in cemented ground can be constructed by excavating a trench, lining the side with several impermeable sheets and backfilling the trench with bentonite cement (Fig. 11.1d). A slurry trench technique may be possible in weak ground, but offers no guarantee of watertightness because it cannot be inspected. Sheet piling has been used for both large (Kitching & Day, 1979) and small lysimeters. Once the monolith is isolated, a trench is dug to expose the piles; the bottom of the piles are trimmed if necessary to form a neat, horizontal base and the joints between them are filled with bitumen or similar compounds. A third technique, forcing a fibreglass cylinder into the ground, has been developed by agriculturalists (Belford, 1979). Cylinders, 0.8m in diameter, 1.5m deep and of 10mm wall thickness have been successfully installed in a variety of soils and weak rocks, using ground anchors for reaction.

An artificial base is needed if the lysimeter sides do not reach an impermeable stratum. Near-horizontal sheet piles have been used for a few large lysimeters, but require specialised equipment. A more common method is to jack a steel plate under the bottom of the monolith from the access shaft or trench. In some cases the monolith has been slid across onto the plate. The sides and base are sealed to each other and most of the access trench backfilled.

11.2.4 Instrumentation and measurements

The minimum instrumentation for a lysimeter site is:

- drainage collector
- daily raingauge
- water table observation borehole.

Other desirable instrumentation includes:

- soil moisture and pressure measuring devices, ie neutron probe and tensiometers
- climate station for potential evapotranspiration estimates
- direct measurement of evapotranspiration
- recording raingauge.

In a humid climate, the minimum measurements would be of daily drainage and rainfall, weekly measurements of water table level. In climates where single rainfall events are important, measurement frequency must be based on events.

11.2.5 Advantages and disadvantages of lysimeters

Lysimeters, as with most other methods of estimating direct recharge, are better suited to humid than semi-arid climates. Careful construction, regular maintenance and frequent observation are needed to ensure that leaks, bypassing flows, faulty drainage or evaporation do not affect the amount of water collected.

Lysimeters are expensive to construct and only give point measurements of recharge. Kitching et al. (1977), set up two lysimeters within 50 metres of each other in a humid zone. Recharge was 159 mm/y in one and 114 mm/y in the other. The fact that the latter recharge value is only 72 % of the former was not explained: does it represent true variability in recharge or errors in measurement? In either case it shows that lysimeter methods are as variable as other point methods.

The act of constructing a lysimeter will disturb the soil and moisture and time is needed for flows to settle back to a natural condition. At the extreme, this could be as long as the time for recharge to flush the lysimeter through; for a 2 m deep unit with a water content of 20% and recharge of 100 mm/y, this would be 4 years. For example, a lysimeter in Cyprus only recorded 5mm of recharge in its first year of operation, although chloride and tritium profiles suggested average recharge was 50 mm/y (Kitching et al., 1980; Edmunds & Walton, 1980).

Lysimeters may be justified where there are a limited number of surface and soil conditions in a catchment, so that the results can be reasonably applied to a large area; see Section 11.7.

In arid and semi-arid areas, because of the variability of precipitation and the importance of rare events, a long period of measurement is needed to estimate average recharge. However, a short record from a lysimeter should provide good quality data to calibrate a soil moisture budgetting model (Section 11.4), for an area with long term meteorological data but with no other method of confirming the choice of model and potential to actual evaporation function.

11.3 Empirical methods

Many attempts have been made to find simple relationships between precipitation and recharge. Once derived by careful study, these are commonly used as 'black boxes', making recharge estimates without further consideration of hydrogeology or whether the results are feasible. The simplest empirical formula takes recharge as a proportion of precipitation:

$$r = f\,p \quad \text{(notation below eqn 11.6)} \qquad 11.3$$

where f will probably vary with terrain and climate. The second level of formula includes a threshold. For example

Mandel & Shiftan (1981) give a formula for recharge in Mediterranean climates of:

$$r = 0.9\ (p - 360) \qquad 450 < p < 650\ \text{mm/y} \qquad 11.4$$

More complex formulae often do not preserve dimensionality, for example the Cheeturvedi formula (Sinha & Sharma, 1988) for recharge in India:

$$r = 50.8\ (p/25.4 - 15)^{0.4} \qquad p > 380\ \text{mm/y} \qquad 11.5$$

or Turc's (1954) formula which includes mean annual temperature:

$$r = p\ [1 - (0.9 + p^2/L^2)^{-0.5}] \qquad 11.6a$$

$$L = 300 + 25T + 0.05\ T^2 \qquad 11.6b$$

where r : annual average recharge rate (mm/y)
p : annual precipitation (mm/y)
f : empirical constant
T : mean annual temperature (°C)

The origin of many formulae are lost in the darkness of history. In general they will have been obtained for a particular basin by correlation of precipitation with estimates of recharge obtained by other methods (water table rise, basin discharge, etc).

There are two issues that must be resolved before an empirical formula can be used; 'How reliable is its derivation?' and 'Can it be transposed to either another catchment or another period in time?'. For example, how accurate were the recharge estimates used to derive the formula; were they checked, for example, by calibration of a groundwater model? When transposing, are conditions in the new catchment (or time period) the same as those in the original - depth to water table, unsaturated zone processes, land use, topographical characteristics, climate and type of rainfall.

An illustration of the low accuracy of simple precipitation-recharge relations can be found in a study of 63 desert catchments in Nevada, USA, carried out by Watson et al. (1976). They correlated recharge (estimated from measurements of groundwater discharge) with precipitation, obtaining results as follows:

Precipitation zone (mm)	Percentage recharge	95% confidence interval (%)
> 510	24	±15
380-510	19	±16
300-380	-0.1	±6
200-300	4	±2
< 200	0	±1

The wide confidence intervals make the coefficients unusable for prediction, despite being derived from a large, carefully assembled database.

Given the difficulties outlined above, the occasions when empirical formulae will be most useful are:

(i) for reconnaissance studies, when high margins of error are acceptable;

(ii) for underexploited catchments, or those where groundwater conditions have little effect on recharge;

(iii) for extrapolation backwards in time within the catchment where the formula was derived;

(iv) when the resulting estimates can be checked and amended if necessary, for example when a groundwater model is to be constructed.

Many empirical formulae have been derived and used, especially in project studies by consultants and others, although few have found their way into the scientific literature. It is not possible to give enough background in this manual on each formula to allow its use, and so none can be recommended.

The use of empirical formulae has similarities to the use of *representative basins*. These well instrumented catchments have been set up in many parts of the world to provide accurate estimates of the components of the water balance, including recharge. The same difficulties of transposition will arise with representative basins as with empirical formulae, but at least the basic data will be of high quality. Setting up representative basins is discussed in a number of manuals and conference proceedings, including Anon (1972-3), Body (1982) and Toebes & Ouryvaev (1970).

11.4 Soil moisture budgetting methods

11.4.1 Introduction

Soil moisture budgetting methods are those that measure or estimate the items on the right hand side of eqn 11.1. At its simplest this leads to a conceptual model like that in Fig. 11.2. Water is held in a soil moisture store; precipitation adds to the store, evapotranspiration depletes it. When full, excess precipitation is routed to groundwater as recharge. The most difficult item to measure is actual evapotranspiration, and in general a conceptual quantity called 'potential evapotranspiration' is defined. A budgetting procedure involving soil moisture is used to convert potential to actual evapotranspiration. There are many variations and refinements on Fig. 11.2 some of which are discussed below, but at this stage it is important to note these points:

- such models are only simple conceptual models of the precipitation-recharge process and may not be correct for your situation;

- the essence of the model is the relation between potential and actual evapotranspiration;

- estimates are for a uniform zone (Section 11.7).

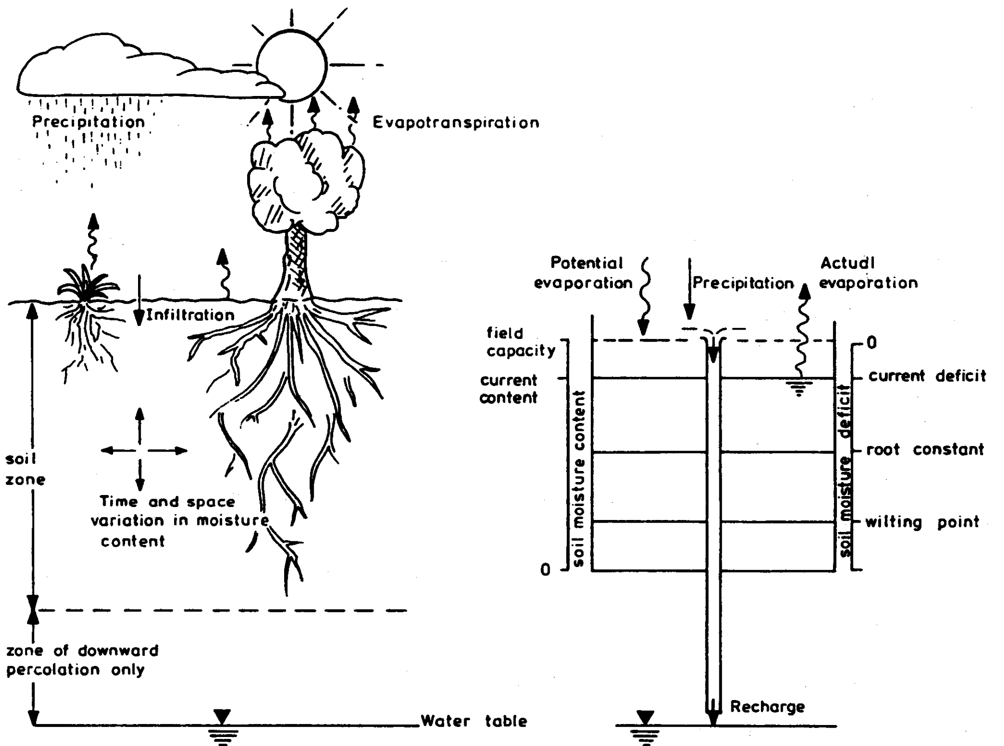


Fig. 11.2 Soil moisture processes and a conceptual soil moisture budgetting procedure

Soil moisture budgetting models were developed for humid climates and have less validity in arid and semi-arid zones. They work best for seasonal patterns of recharge, well developed soils which do not dry completely, when potential and actual evapotranspiration are of similar sizes, and with precipitation that is widespread and relatively uniform. These conditions do not apply in arid and semi-arid zones where these models normally underestimate recharge, often giving zero values.

The following rule of thumb may help in deciding whether soil moisture budgetting models will be applicable to a given terrain:

Whole year: $p + i > 500$ (11.7a)

Wet season: $et_p < 1.5\ (p + i)$ (11.7b)

Dry season: $et_p < 3\ (p + i)$ (11.7c)

et_p, et_a and $(p+i)$ similar in area where et_p data derived and area where model is applied (11.7d)

where et_p : potential evapotranspiration (mm/y)
et_a : actual evapotranspiration (mm/y)
p : average precipitation (mm/y)
i : irrigation water applied (mm/y)

It may be possible to apply the methods for a wet season only on the basis that little recharge occurs in the dry season.

Fig. 11.2 does not describe how soil moisture behaves; there can be vertical water flows (either up or down) when a soil moisture deficit exists. Nor does it describe the recharge process, which may be dominated by fissures, root channels or topographic depressions. Many models used in real situations need empirical adjustments to make them match field conditions. This section outlines the classical, humid zone models that are in use, and lists some of the empirical adjustments used for various studies.

11.4.2 Evapotranspiration

Good data on actual evapotranspiration is equally important as good precipitation data. Unfortunately actual evapotranspiration is rarely measured except in research projects when the sophisticated *eddy correlation* technique is used; a description is given by Dyer & Maher (1965). This is an impractical technique for most project studies, and actual evapotranspiration must be estimated from standard meteorological measurements. As mentioned above, this is usually done through a conceptual quantity called *potential evapotranspiration* or *reference crop evapotranspiration*. These are intended to be measures of the energy available for evaporating and transpiring water.

A number of formulae have been devised for evapotranspiration, of which the Penman-Monteith (Monteith, 1965, 1981) is generally considered the best for actual evapotranspiration in humid climates. It is written

$$et_a = \frac{s\ H\ +\ d\ c\ (e_a - e_d)/r_a}{l\ [s + g\ (1 + r_s/r_a)]} \qquad 11.8$$

(Notation is below eqn 11.10.)

It is rarely used because of the difficulty of estimating the aerodynamic and stomatal resistance parameters; it is however recommended for use whenever possible.

The Penman (1948, 1949) formula is in common use among hydrologists and agronomists to estimate a reference evapotranspiration from short, green, well watered turf. It is written

$$et_t = \frac{s\,H + g\,E_a}{l\,(s + g)} \qquad 11.9a$$

where

$$E_a = a_3\,(\,a_1 + a_2\,u_2\,)(e_a - e_d) \qquad 11.9b$$

(Notation is below eqn 11.10.)

In both formulae, evaporation depends upon the vapour pressure deficit, which in turn depends upon the amount of evaporation. Therefore they do not calculate evaporation from independant variables, and are only strictly valid for description rather than prediction. For example it would be incorrect to take data from a climate station in a large, well irrigated area of crops and apply the resulting evaporation estimate to natural semi-arid vegetation; this is one reason for rule 11.7d above. De Bruin (1987) discusses this and other problems in an excellent review of evaporation in arid and semi-arid regions.

Despite the theoretical difficulties, the Penman and Penman-Monteith formulae are, and will continue to be, used to estimate actual and potential evapotranspiration. Actual evapotranspiration will continue to be estimated from potential by the models described in this section. Manuals on the use of Penman should be consulted including WMO (1966), FAO (1976) and FAO (1977).

Other evaporation formulae include:

- Thornthwaite (1948) using only monthly mean temperature, although a later version used daily temperatures;

- Turc (1954) using annual precipitation and mean temperature (see eqn 11.6);

- Priestley and Taylor (1972), a simplification of Penman-Monteith;

- Thom and Oliver (1977), a revision of Penman-Monteith.

None of these are recommended. Most use less data than Penman or Penman-Monteith, and none have been researched as thoroughly as Penman. They are unlikely to provide accurate enough estimates for recharge estimation, especially in arid and semi-arid areas.

Recently the Dutch have gone over to the use of a simple formula by Makkink that only requires global radiation and temperature (Hooghart, 1987)

$$et_r = 0.65\ s\ G/[l\ (s + g)] \qquad 11.10$$

where
- et_a : actual evapotranspiration (kg/m²/s)
- et_t : evapotranspiration for short, well watered, green turf; Penman potential evapotranspiration (kg/m²/s)
- et_r : reference crop evapotranspiration (kg/m²/s)
- a_1, a_2, a_3 : constants
- c : specific heat of air (J/kg/K)
- d : density of air (kg/m³)
- e_a : saturated vapour pressure at air temperature (mbar)
- e_d : vapour pressure at screen height (mbar)
- g : psychrometric constant (mbar/K)
- G : global radiation at edge of atmosphere (w/m²)
- H : available energy = net radiation - soil heat flux - heat storage in vegetation (W/m²)
- l : latent heat of vaporisation (J/kg)
- r_a : aerodynamic resistance (s/m)
- r_s : stomatal resistance (s/m)
- s : slope of saturated vapour pressure curve (mbar/K)

It estimates a reference crop evapotranspiration which is converted to actual evapotranspiration by published crop factors. These are an essential part of the procedure, and would need to be evaluated for other crops and climates to those in The Netherlands.

11.4.3 Penman-Grindley model

The Penman-Grindley model (Penman 1950; Grindley 1967, 1969) is the simplest and most widely used soil moisture budgetting model. It was originally developed to estimate soil moisture deficit and actual evaporation, recharge estimates being a by-product. The soil moisture accounting is common to many models:

$$\begin{aligned} psmd_{i+1} &= smd_i + ae_i - p_i \\ r_i &= -\ psmd_{i+1} \quad \text{when } psmd_{i+1} < 0 \\ smd_{i+1} &= psmd_{i+1} - r_i \end{aligned} \qquad 11.11a$$

Actual evapotranspiration is derived from potential as follows:

$$\begin{aligned} ae_i &= pe_i && \text{when } smd_i < C \text{ or } p_i \geq pe_i \\ ae_i &= p_i + F\ (pe_i - p_i) && \\ & && \text{when } D > smd_i \geq C \text{ and } p_i < pe_i \\ ae_i &= p_i && \text{when } smd_i = D \text{ and } p_i < pe_i \end{aligned} \qquad 11.11b$$

where
- smd_i : soil moisture deficit at the start of day or time period i
- ae_i : actual evapotranspiration during day i (L)
- pe_i : potential evapotranspiration during day i (L)

p_i : precipitation during day i (L)
r_i : recharge during day i (L)
$psmd_i$: intermediate variable (L)
C : root constant (L)
D : wilting point (L)
F : empirical constant

The shape of this function is shown in Fig. 11.3, which shows the three parameters (C,D,F) of the model which must be calibrated or estimated. F is an empirical constant relating actual evapotranspiration to potential when deficits are greater than the the root constant. All three are related to the vegetation cover and, as a second order effect, to soil characteristics. Table 11.2 gives the values of C and D usually used in the UK, where a value for F of 10% is adopted.

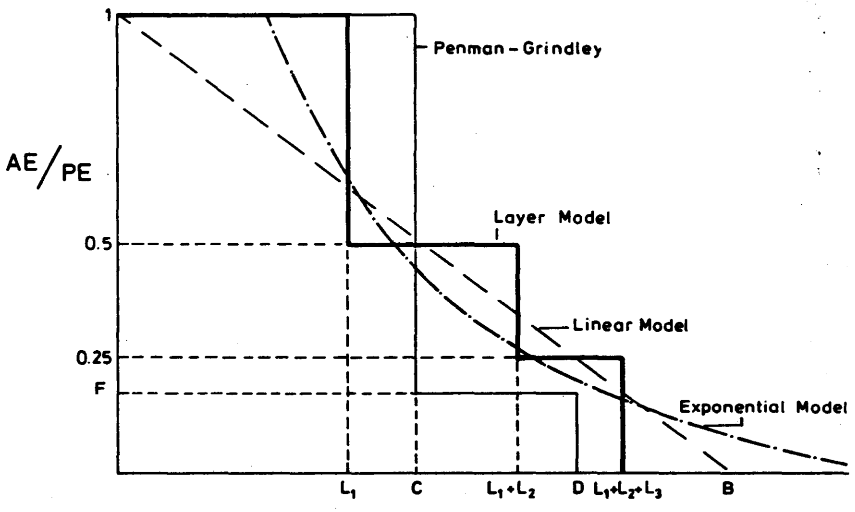


Fig. 11.3 The form of actual -vs- potential evapotranspiration relationship used in various soil moisture budgetting models. Based on Calder et al., 1983

11.4.4 Other models

A number of other models have been used. They mainly differ from the Penman-Grindley model in the shape of the actual-potential evapotranspiration relationship, as shown in Fig. 11.3. They include:

- layer models, such as those described by Calder et al. (1983), Palmer (1965) and Stuff & Dale (1978);

- a linear model (Calder et al., 1983);

- exponential models (Calder et al., 1983; Johannson, 1987).

It will usually be found that the difference between models will be less important than the accuracy of precipitation, potential evapotranspiration, irrigation and cropping data.

Table 11.2 Monthly root constant (C) and wilting point (D) values for the Penman-Grindley model in the UK (mm)

Month		Crop type (see notes) 1	2	3	4	5	6	7	8	9	10	11	12	13	14
Jan & Feb	C	25	25	25	25	25	25	25	25	25	25	56	76	13	203
	D	25	25	25	25	25	25	25	25	25	25	102	127	51	254
Mar	C	56	56	56	25	25	56	25	25	25	25	56	76	13	203
	D	102	102	102	25	25	102	25	25	25	25	102	127	51	254
Apr	C	76	76	76	76	56	56	56	25	25	25	56	76	13	203
	D	127	127	127	102	102	102	102	25	25	25	56	76	13	203
May	C	97	97	97	56	56	56	56	56	25	25	56	76	13	203
	D	152	152	152	102	102	102	102	102	25	25	102	127	51	254
Jun & Jul	C	140	140	140	76	76	25	56	56	56	25	56	76	13	203
	D	203	203	203	127	127	25	102	102	102	25	102	127	51	254
Aug	C	140	140	25	97	97	25	25	56	56	25	56	76	13	203
	D	203	203	25	152	152	25	25	102	102	25	102	127	51	254
Sept	C	140	25	25	97	25	25	25	25	56	25	56	76	13	203
	D	203	25	25	152	25	25	25	25	102	25	102	127	51	254
Oct	C	25	25	25	97	25	25	25	25	56	25	56	76	13	203
	D	25	25	25	152	25	25	25	25	102	25	102	127	51	254
Nov & Dec	C	25	25	25	25	25	25	25	25	25	25	56	76	13	203
	D	25	25	25	25	25	25	25	25	25	25	102	127	51	254

Notes

1. Values originally quoted in inches (1 in = 25.4 mm) and rounded to nearest mm for this table.
2. Valid for England and Wales only.
3. Crop types are:
   - 1 cereals, Sept harvest
   - 2 cereals, Aug harvest
   - 3 cereals, July harvest
   - 4 potatoes, Sept harvest
   - 5 potatoes, May harvest
   - 6 vegtables, May harvest
   - 7 vegtables, July harvest
   - 8 vegtables, Aug harvest
   - 9 vegtables, Oct harvest
   - 10 bare fallow
   - 11 temporary grass
   - 12 permanent grass
   - 13 rough grazing
   - 14 woodland
   - 15 riparian (not shown) - C and D effectively infinite
4. Based upon Grindley (1969).

11.4.5 Runoff

On less permeable soils a significant proportion of precipitation may become runoff. This is usually taken as an empirically derived proportion

$$q = f\ p \qquad 11.12a$$

or a threshold may be included

$$q = f\ (p - p_t) \quad \text{if } p>p_t \qquad 11.12b$$
$$q = 0 \quad \text{if } p<p_t$$

or the runoff proportion may be related to the soil moisture deficit

$$q = f\ p\ (1 - smd/C) \qquad 11.12c$$

where
- q : runoff in time period (L)
- f : empirical factor (<1)
- p : precipitation (L)
- p_t : threshold precipitation below which no runoff occurs (L)
- smd : soil moisture deficit (L)
- C : empirical constant

The factors f, p_t and C are found by calibration against measured runoff, perhaps from an experimental catchment. There may be difficulty is deciding what part of the total runoff is due to the soil. In a model which is lumped over an area, part of the runoff may come from particular zones, for example the riparian zone where the water table is shallow. The approach described in this paragraph has similarities to catchment modelling (FAO, 1978).

An alternative method of estimating runoff is to measure the soil infiltration capacity and compare this with short duration rainfall rates; data from recording raingauges and measurements of infiltration rate against time are needed. The analysis is lengthy and complex if a representative set of storm profiles and sequences is to be analysed.

11.4.6 Other empirical adjustments

Many authors alter their models to allow for crop ripening and removal. Two methods are (i) to alter the actual-potential evapotranspiration relationship, for example by reducing the root constant C in the Penman-Grindley model or (ii) changing the potential evapotranspiration estimates to reflect the state of the vegetation.

Recharge is seen to occur in some aquifers even when there is a soil moisture deficit. Rushton & Ward (1979) explored a number of ways of allowing such recharge in a soil moisture budgetting model. For a limestone aquifer in Eastern England, they finally chose a constant proportion of precipitation when it exceeded a threshold. The remaining precipitation entered a conventional Penman-Grindley model and could give rise to additional recharge. Various other empirical estimates of such rapid or bypassing recharge have been used; a method must be chosen and calibrated to suit local conditions.

11.4.7 Time steps

Numerous authors (eg Howard & Lloyd, 1979) have pointed out that the time-step used in soil moisture models is critical. Longer time-steps, with the same parameters, lead to lower or zero recharge estimates. All recent work recommends a daily time step for humid zones. Intervals of less than a day, eg ones for both day and night, or storm based intervals, may be needed in arid and semi-arid areas, that is if the methods can be made to work at all.

11.4.8 Calibration of models

There is no universally correct soil moisture budgetting model. Numerous authors have shown how different models, and different parameters, can significantly alter the recharge estimate (eg Howard & Lloyd, 1979; Alley, 1984; Calder et al., 1983; Rushton & Ward, 1979). For any situation, a model should be chosen based on a conceptual model of the local recharge processes. This model should then be calibrated by one of the following methods, for point estimates:

(i) against a lysimeter,

(ii) against soil moisture measured by a neutron tube or tensiometer;

and for areal estimates:

(iii) against other estimates of recharge, eg from groundwater flow modelling,

(iv) against a catchment water balance.

11.5 Darcian approaches

11.5.1 Introduction

The flow of water in the unsaturated zone is governed by Darcy's Law, with the difference from fully saturated flow that hydraulic conductivity varies with moisture content. Moisture content and hydraulic conductivity also vary with pressure in the unsaturated zone. This section presents a brief review of the theory, and then describes how knowledge of these three parameters can be used to estimate the vertical flow of recharge, either by numerical modelling or by interpretation of field measurements.

11.5.2 Theory of flow in the partially saturated zone

Water, whether in the fully or partially saturated zones, has a total potential, h, which is the sum of two components; the gravitational component, z, and the pressure component, p:

$$h = z + p \qquad 11.13$$

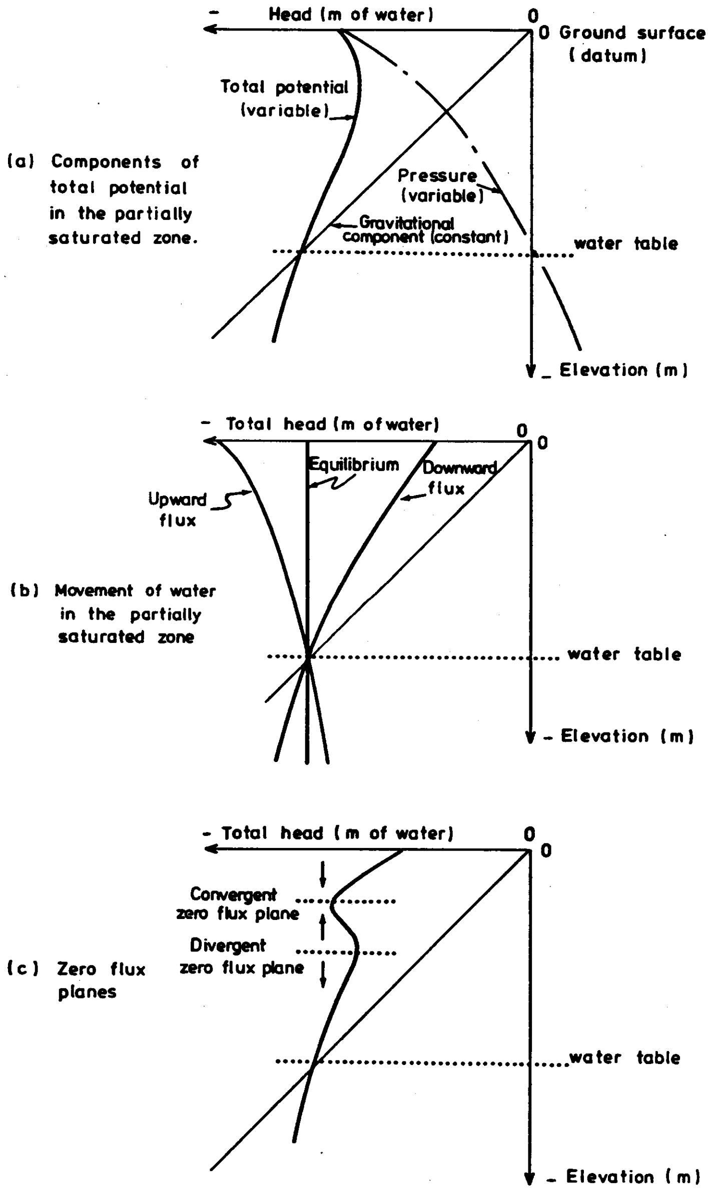


Fig. 11.4 Head distributions and water movement in the unsaturated zone

It is convenient to work in terms of head of water, in which case z is the elevation relative to datum. For unsaturated flow, the ground surface is often taken as the datum. The variation of z with depth is always the same and is shown in Fig. 11.4a as the gravitational component.

In the unsaturated zone, the pressure arises from the capillary forces that hold water in the smaller pores; it is always negative, that is below atmospheric pressure. Below the water table, pressure is a combination of hydrostatic forces (the weight of water above) and flow forces; and is always positive. An idealised curve is shown in Fig. 11.4a.

Darcy's Law states that the seepage velocity is related to hydraulic conductivity and gradient of total head by:

$$q = -k \frac{\delta h}{\delta x} \qquad 11.14$$

(Notation for this section below eqn 11.17.)

Flow is usually vertical in the unsaturated zone, and hydraulic conductivity depends on moisture content, so Darcy's law is rewritten:

$$q = -k_\theta \frac{\delta h}{\delta z}, \text{ where } k_\theta = f(k,\theta) \qquad 11.15$$

Thus flows depend on the profile of total potential, which is obtained by summing the gravitational and pressure components as shown in Fig. 11.4a. When total potential is constant, the profile is in equilibrium and no flow is occurring (Fig. 11.4b). When the gradient is upwards, flow is upwards, perhaps to satisfy evaporative demand in the root zone (Fig. 11.4b). Conversely, when the gradient of total potential is down, the flow is downwards (Fig. 11.4b). Both upward and downward flow can occur in the same profile, in response to time variant inputs (Fig. 11.4c). The zones of upward and downward flow are separated by a "zero flux plane", so called because there is no flow across it.

In transient situations, water is taken up or released from storage by changing the moisture content or degree of saturation. A differential equation to describe this process can be obtained by combining Darcy's Law with a mass conservation equation. The result is Richards' equation (1931):

$$\frac{\delta \theta}{\delta t} = \frac{\delta}{\delta z}\left[k_\theta \frac{\delta h}{\delta z} \right] \qquad 11.16$$

This equation is sometimes written in terms of pressure only or moisture content only (Freeze, 1969). There may be a sink term required to account for, as an example, evapotranspirative moisture uptake by roots. This could be included as $S(\theta,z,t)$:

$$\frac{\delta\theta}{\delta t} = \frac{\delta}{\delta z}\left[k_\theta \frac{\delta h}{\delta z}\right] - S(\theta,z,t) \qquad 11.17$$

where
- q : seepage velocity (L^3/T)
- k : hydraulic conductivity (L/T)
- k_θ : unsaturated hydraulic conductivity (L/T)
- θ : moisture content
- h : hydraulic head (L)
- S : rate of outflow of moisture ($1/T$)

A major difficulty with flow in the unsaturated zone is the nature of the relationships of hydraulic conductivity and moisture content to pressure. Both relationships have hysteresis, that is are different for wetting and drying (Fig. 11.5).

2'

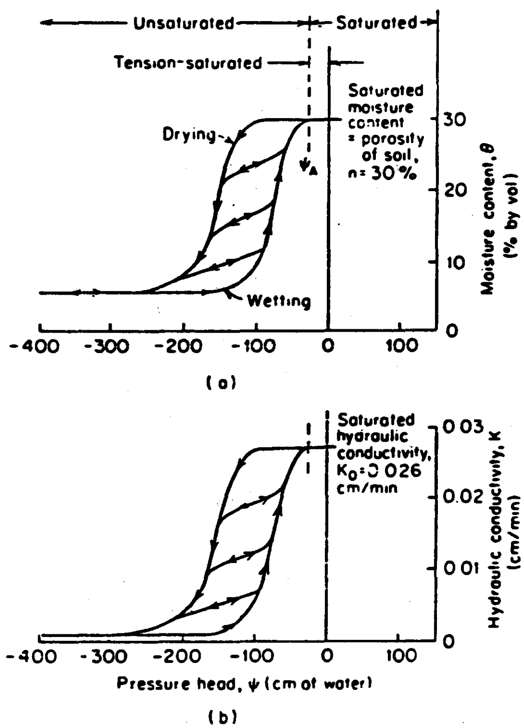


Fig. 11.5 Characteristic curves relating hydraulic conductivity, moisture content and pressure head for a naturally occurring sandy soil (after Freeze & Cherry, 1979, Fig. 2.13)

11.5.3 Numerical solution of Richards' equation

Freeze (1969) wrote one of the first finite difference computer models and gives a full description of his implicit solution procedure. He did not make any restrictive assumptions, and allowed for hysteresis and a dynamic water table. A later paper compared the model against a laboratory column experiment and good agreement was claimed (Freeze & Banner, 1970).

Other authors who have developed numerical models for unsaturated flow include Krishnamurthi et al. (1977), Kafri & Ben Asher (1978), Watson (1981), Jansson & Halldin (1979, 1980), Davidson (1985), and Morel-Seytoux & Billica (1985). Few models have been applied to real field problems, and even fewer have been applied to arid and semi-arid areas. Jansson and Halldin's model has been extensively used in Sweden, where a database of soil properties has been built up; see for example Johansson (1986, 1987). The Dutch have a standard model for areas in their country with shallow water tables (TNO, 1981).

11.5.4 Difficulties when modelling unsaturated flow

This section does not address the problems of developing a suitable computer code. Methods have been published, and codes for all types of numerical model are available at low cost through:

International Groundwater Modelling Center,
Butler University,
Indianapolis,
Indiana 48208,
USA

The discussion here concentrates on the difficulties of obtaining realistic results from a computer code. These fall into three types; (i) developing a complete conceptual model of the processes occurring in the field, (ii) obtaining reliable field data on soil properties, (iii) calibrating the model.

Conceptual model. The studies mentioned above all assume that flow in the unsaturated zone is vertical. This is true of most such modelling studies in the literature; many also assume that the same vertical flows occur everywhere in the catchment. This simplistic model is generally invalid for two reasons. Firstly, lenses and layers of lower permeability materials are common. These can cause perched water tables to develop and induce horizontal movement of infiltrating water. Secondly, surface topography may be a major influence on infiltration, leading to large variations in recharge across areas of uniform soils. Freeze & Banner (1970), among others, noted that recharge was focussed in depressions even in a high permeability gravel aquifer. The numerical models outlined above do not usually incorporate features observed in the field such as (a) drying cracks, (b) swelling of clays on wetting, (c) macropores and root channels, all of which dominate recharge processes in many soils.

Data on soil properties. The three variables k, θ, and p are interrelated, and two of the three relationships are needed. They are all non-linear and often show hysteresis (Fig. 11.5). These relationships must be measured for each significant layer in order to create a realistic numerical model. Methods are discussed by Brooks & Corey (1964), Bouwer & Jackson (1974), and in soil physics textbooks.

Calibration of the model. Most models have been used for studies of hypothetical soils, for example in parametric studies. Before such models can be used to estimate recharge, they must be calibrated. This can be site specific or regional. Site specific calibration would be against field measurements of recharge, eg from a lysimeter or from moisture content or pressure in an instrumented profile. Site specific calibration leaves the problem of extrapolation in space, but should yield a realistic model for that site. Regional calibration, for example against a groundwater model or water table fluctuations, must force the unsaturated flow model to use "averaged" properties. These may be unrealistic, particularly when extrapolated in time.

11.5.5 Use of field data from the unsaturated zone

There are three situations where measurements of pressure and moisture content in the unsaturated zone can potentially be used to estimate recharge:

(i) When there is no input to the soil profile at the surface, and the profile is draining to the water table or to evapotranspiration,

(ii) When there is input, but it is insufficient to saturate the soil,

(iii) When there is sufficient surface input to saturate the profile.

No input. When there is no precipitation input to a soil profile, evaporation rapidly lowers the moisture content and pressure near the surface. A zero flux plane develops and can be detected by tension measurements. Below the zero flux plane, water drains downward to become recharge (Fig. 11.6). The change in storage in each zone is the amount of evapotranspiration or recharge:

$$et_a = \sum_{i=ground}^{zfp} [\theta_i(t_1) - \theta_i(t_2)]\ l_i \qquad 11.18a$$

$$r = \sum_{i=zfp}^{wt} [\theta_i(t_1) - \theta_i(t_2)]\ l_i \qquad 11.18b$$

where et_a : actual evapotranspiration (L)

r : recharge (L)
θ_i : soil moisture content at i
t_1 : start of time interval (T)
t_2 : end of time interval (T)
l_i : length of sampling interval (L)
zfp : zero flux plane
wt : water table

The zero flux plane and water table will move over the time interval (t_1, t_2) as shown in Fig. 11.6. The area A represents the volume evaporated, the area B represents the volume draining to the water table. Measurements of pressure and soil moisture have been used to estimate recharge in this manner. For example, Wellings (1984) gives examples for chalk in southern England where there is continuous drainage (recharge) throughout the summer despite the potential evapotranspiration greatly exceeding precipitation in these months. Cooper (1980) gives an example in eastern England where 43% of annual drainage occurred when there was a zero flux plane.

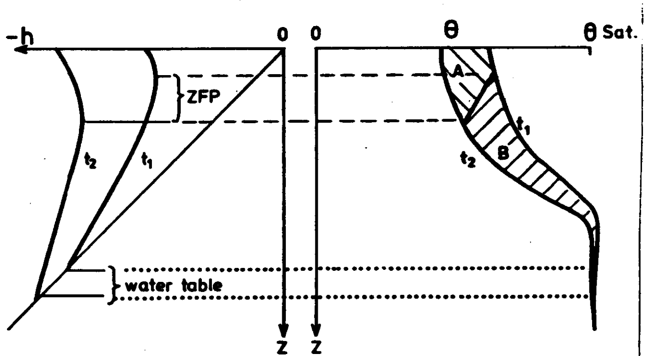


Fig. 11.6 Evaporative (A) and drainage (B) fluxes from a profile with no surface input

Small input. The inputs of water to a soil profile usually vary in time, and so there are transient changes in the soil moisture and pressure. Field measurements of these variables were used by Steenhuis et al. (1985) on a plot in Long Island, USA, which has a rainfall of about 2000 mm/y. However, field measurements cannot normally be made accurately enough to calculate unsaturated, transient flows, especially in arid and semi-arid areas, and can only be used in cases where steady state can be assumed.

Steady state conditions will never apply near the surface nor in the root zone. However, a thick unsaturated zone will tend to dampen and coalesce seasonal (or storm) pulses of recharge and so may reach an approximate steady state at depth. In this condition the unsaturated form of Darcy's Law (eqn 11.15) can be applied.

It is generally assumed that pressure is constant with depth under these steady state conditions, so that the hydraulic gradient is 1 and entirely due to the gravitational component of head. In this case:

$$r = k_\theta \qquad 11.19$$

where r : recharge rate (L/T)
k_θ : unsaturated hydraulic conductivity (L/T)

and both r and k_θ should be the same at all depths. This implies that layers which have different saturated conductivities will be at different saturations to make their unsaturated conductivities equal.

The field data requirements are (i) measurements of θ and p at several depths, (ii) k_θ - θ relationships for the material at each depth. The values of pressure are used to check that gradients are negligible, the θ values are used to interpolate k_θ from k_θ - θ relationships.

Sammis et al. (1982) describe briefly how the method was applied to a site in Arizona where the water table was 42m below the ground. Push tube samples were taken every 3m during drilling of a borehole. Pressure was found by inserting a tensiometer into the samples; moisture contents were measured by the gravimetric method. They give few details of their method for unsaturated conductivity, refering only to papers by Millington & Quirk (1959, 1960, 1961). Their results are not impressive with k_θ estimates of 12, 790, 0.005, 226, 0.2 and 23 mm/y for various depths; under the method's assumptions, all these values should be the same.

Large input. If the infiltration rate is high enough, the ground will become saturated. Water will move downwards under gravitational forces only, that is the hydraulic gradient dh/dz = 1, provided that ponding on the surface is only slight so that pressure does not rise significantly above atmospheric. In this restricted circumstance:

$$r = k_{sat} \qquad 11.20$$

where r : recharge rate (L/T)
k_{sat} : saturated hydraulic conductivity (L/T)

that is infiltration rates can be estimated from saturated hydraulic conductivity (eg Dreiss & Anderson, 1985). Although appealingly simple, this method has limited usefulness except when continuous applications of water are made, such as in irrigation schemes or river beds. Recharge will be controlled by the lowest permeability layer, which may not be at the surface.

11.5.6 Darcy's law in the saturated zone

Measurements below the water table avoid the difficulties of determining the k-p-θ relationships, but introduces the possibility of flow in the horizontal as well as vertical plane. However, there are situations in which flow beneath the water table is predominately vertical. Rehm et al. (1982)

discuss a case in North Dakota, USA, where a layered aquifer has one high conductivity layer, in which the major horizontal flows occur, overlain by a lower conductivity layer containing the water table. In this case, flows in the upper layer will be predominately vertical, and can be estimated from measurements of (vertical) hydraulic conductivity and heads from a vertical array of piezometers. The flows may vary with time, and so regular or, preferably, continuous measurements of heads are needed. The flows calculated are not equal to recharge at any instant, because of storage effects as the water table rises and falls. However, summing the flows over a year or longer will give a good estimate of total recharge.

11.6 Tracer techniques

11.6.1 Introduction

Tracers (both isotopic and chemical) have been widely used in estimating arid and semi-arid recharge. They can be grouped into:

- environmental tracers, ie those that are already present in the geosphere,

- tracers applied by the researcher.

Table 11.3 Principal applied and environmental tracers

Tracer	Conservative	Applied/ Environmental	Half life
^{2}H, D	N	E	stable
^{3}H, T*	N	A & E	12.43 y
^{13}C	Y	E	stable
^{14}C*	Y	E	5730 y
^{18}O	N	E	stable
Cl	Y	A & E	stable
^{36}Cl*	Y	E	300 000 y
Br	Y	A	stable
^{51}Cr	Y	A	27.8 d
^{58}Co	Y	A	77 d
^{60}Co	Y	A	5.3 y
^{131}I	Y	A	8.04 d

* Natural and manmade (nuclear bomb) sources

The principal tracers are reviewed in Table 11.3. Among the environmental isotopes, ^{18}O and ^{2}H (Deuterium) are affected if water is evaporated. The isotope fractionation processes, occurring between liquid and vapour phases during evaporation, impose changes in the isotope composition of the water remaining in the soil moisture, which would need to be accounted for in any quantitative method. On the other hand, such changes often provide a useful label on the water to study the origin and processes of recharge. Transpiration by plants does not cause such fractionation. Tritium, the most widely used environmental tracer for recharge studies, is not usually considered to be affected. Conservative, ie not evaporated, tracers are concentrated by evaporation and so maintain their accuracy at low rates of recharge.

The detailed methodology and the theoretical basis for the use of stable isotopes are described in *Stable isotope hydrology* (IAEA, 1981).

Environmental tracers can be assumed to be applied evenly over the whole land surface. Their movement in the unsaturated zone therefore is only in one dimension, with no lateral diffusion or dispersion. Applied tracers on the other hand are applied at a point or over a small area and so can disperse laterally as well as move vertically. This means that checks on the mass balance of an applied tracer are much more difficult to achieve.

There are several ways that tracers can be used:

(i) signature methods, in which particular parcels of water are labelled and traced. These methods have been used in three different situations:

  (a) environmental tracers in the unsaturated zone,

  (b) applied tracers in the unsaturated zone,

  (c) environmental tracers below the water table;

(ii) throughput methods, when fluxes of tracer and water are calculated in the unsaturated zone, usually for environmental tracers;

(iii) turnover or transit time calculations, which are used for whole aquifers, usually with environmental tracers. These are explained in Chapter 16.

The signature and throughput methods generally assume a single flow system and piston displacement in the unsaturated zone. If there are two flow systems, recharges can bypass the matrix of the unsaturated zone, for example by flowing in fractures, down rootways or other macropores. All of these phenomena have been observed in arid and semi-arid areas and may invalidate the methods. Some workers have interpreted their field data in terms of two flow systems; an example is discussed below in Section 11.6.3.

11.6.2 Signature methods

General principles. Infiltrating water may be labelled (in time) by a particularly noticeable input of tracer. This may either be applied artificially, or be an environmental tracer, for example from an atmospheric nuclear explosion. This labelled water may be identified at depth in the soil profile at future dates, so providing information on water movement in the unsaturated zone.

Once the water at a particular depth has been dated, two methods are available to calculate average recharge. At a time of maximum soil moisture deficit, the total water content above the dated depth is the total recharge over the intervening years, ie:

$$r = 1/n \int_0^d \theta_z \, dz \qquad 11.21a$$

(The notation is below eqn 11.25.)

In practice a summation of the sampling intervals would be used in place of eqn 11.21a:

$$r = 1/m \sum_{i=1}^{m} \theta_i \; l_i \qquad 11.21b$$

The second method is a mass balance of tracer, and is only applied to environmental tracers, or those applied over a large enough area to prevent lateral dispersion. The amount above the dated layer is measured:

$$T_p = \sum_{i=1}^{m} T_i \; \theta_i \; l_i \qquad 11.22$$

and equated with the input to recharge after allowing for radioactive decay:

$$T_p = \sum_{j=1}^{n} r_j \; T_j \; \exp(d_j) \qquad 11.23$$

Some workers estimate the relative sizes of r_i using rainfall and potential evapotranspiration data to define annual weightings:

$$r_j = w_j \; r, \quad \Sigma \; w_j = n \qquad 11.24$$

Combining eqns 11.22-11.24 gives average recharge as:

$$r = T_p \; / \; [\sum_{j=1}^{n} w_j \; T_j \; \exp(d_j)] \qquad 11.25$$

where
- n : number of years between signature and sampling (T)
- r : mean annual recharge (L/T)
- r_j : recharge in year j (L)
- d : depth of dated water (L)
- θ_i : moisture content in sampling interval i
- l_i : length of sampling interval number i (L)
- T_i : tracer concentration in sampling interval i (M/L^3)
- T_p : tracer mass in profile (M)
- T_j : tracer concentration in rainfall in year j (M/L^3)
- $\exp(d_j)$: decay of tracer input since year j
- n : number of years
- m : number of sampling intervals

w_j : ratio of recharge in year j to mean recharge

Other workers have found that assuming equal recharge every year (ie $r = r_j$) is sufficiently accurate (Allison & Hughes, 1978). Other examples of the use of environmental tracers are given by Edmunds & Wright (1979) and Lloyd et al. (1981).

For a conservative tracer, average recharge is estimated from eqn 11.21b. A balance of tracer can then be used to calculate the fraction of precipitation that becomes recharge.

Environmental tracers in the unsaturated zone. The most commonly used environmental tracer for this method has been tritium which, as well as being produced naturally in the atmosphere, was introduced into the atmosphere in large amounts in 1952 by nuclear tests. There was a significant peak, particularly in the northern hemisphere, in 1963/4 (Fig. 11.7). Atmospheric concentrations were always much less in the southern hemisphere and, as atmospheric testing stopped in 1963, tritium levels are now falling everywhere. More sensitive analytical techniques have extended the usefulness of environmental tritium for some time, but it is now rarely of use for signature methods.

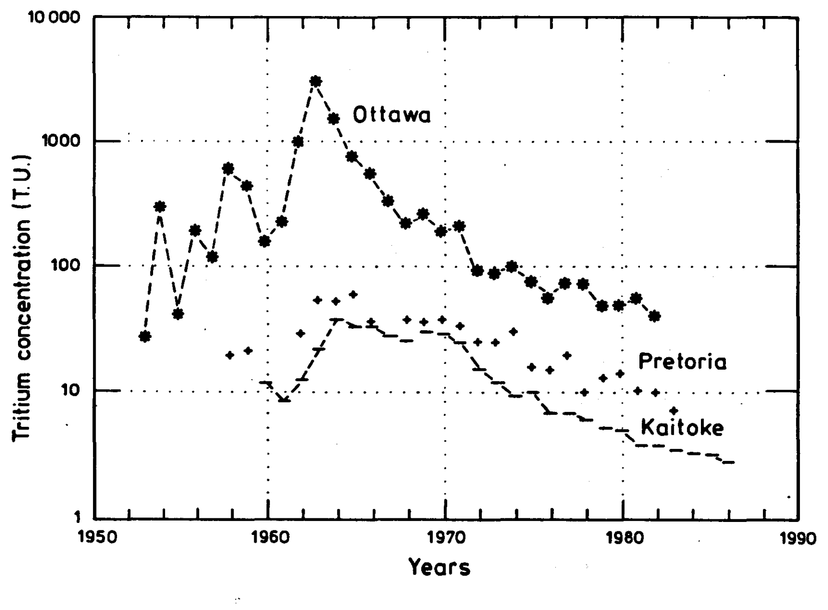


Fig. 11.7 Tritium concentrations in rainfall (data courtesy of IAEA)

Oxygen-18 (^{18}O) and Deuterium (^{2}H) are fractionated during precipitation and evaporation processes, a property which has so far prevented their use in any tracer balance method of estimating recharge. They have shown useful signatures in some situations, but these have been in temperate, high recharge areas. For example Thoma et al. (1979) found seasonal markers of deuterium in sand dunes; Bath et al. (1982) and Saxena & Dressie (1984) sometimes found cyclic profiles of ^{18}O and ^{2}H corresponding to seasonal rainfall and recharge; snowmelt often contains a distinctive $^{18}O/^{2}H$ signature. However Allison (1988) reports that such seasonal peaks are not seen in arid and semi-arid areas because diffusion will redistribute the peaks over the very short interval between them. He also reports that the displacement of ^{18}O and ^{2}H isotopic ratios from the local meteoric line may be related to recharge rates, but that a technique needs further development.

Applied tracers in the unsaturated zone. The signature method is mainly used with applied tracers. Radio-isotopes are usually used as they can be detected at low levels, thus they can be introduced in small quantities without large disturbances to soil or its moisture content. The tracer is introduced below the lowest depth of the zero flux plane, that is below the region where upward flow or evapotranspiration losses can occur (Fig. 11.4). The method is widely used in India, and Fig. 11.8 shows a typical site layout for tritium and Cobalt-60 (^{60}Co) injection.

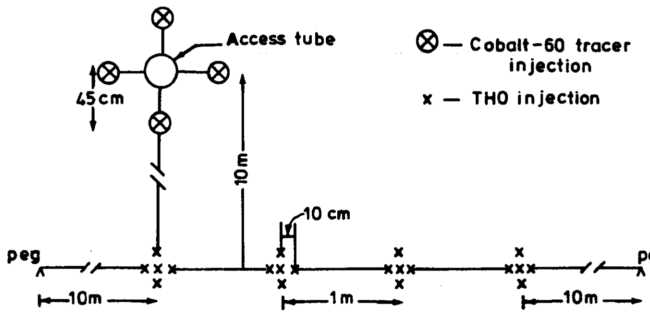


Fig. 11.8 Layout of injection points for Tritium and Cobalt-60 (after Chandrasekharan et al., 1988)

The tracer position is determined at a future date, for example at the end of the wet season or one or more years later. The tracer may be detected by drilling a cored borehole and analysing the porewaters (^{3}H), or by non-destructive measurements of radiation from an adjacent open borehole (^{60}Co). Inevitably the injected tracer will have dispersed (Fig. 11.9). Conventionally the centre of gravity of the tracer profile is taken to show how much displacement there has been; the moisture contained in the soil between injection point and centre of gravity is equivalent to recharge over the time period. Examples of the use of applied tracers are given by Athavale et al. (1980), Sharma et al. (1985), Chandrasekharan et al. (1988) and Athavale & Rangarajan (1988).

Environmental tracers below the water table. Because the peak of bomb tritium has moved out of the unsaturated zone in most parts of the world, there have been recent attempts to use its distribution below the water table to estimate recharge. The method relies on obtaining a profile in an area where groundwater only moves vertically. Profiles have been taken on groundwater divides and in the centre of recharge mounds. The formulae are as above. In all three published case studies of alluvial aquifers (Knott & Olimpio, 1986; Larson et al., 1987; Delcore & Larson, 1987), less tritium has been found in the profile than would be expected from recorded inputs, even after allowance for decay, seasonally varying inputs and dispersion. This suggests that there is an inconsistency in the conceptual model of the method or aquifer, for example that there are multiple flow pathways in granular materials, or that it has not been possible to estimate inputs correctly. The method needs further research.

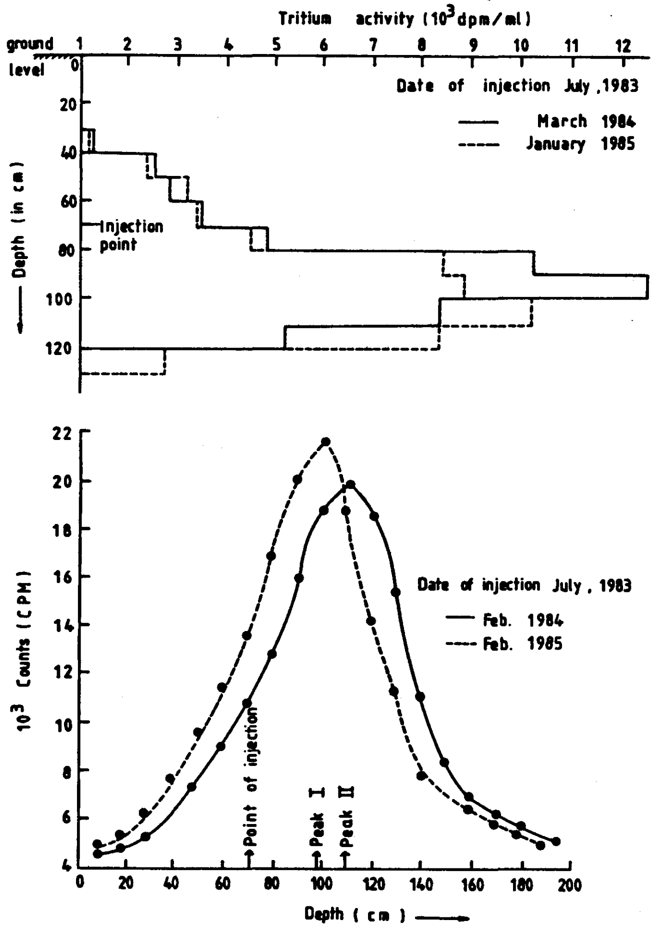


Fig. 11.9 Tracer profiles for Tritium and Cobalt-60 at a site in Jodhpur (after Chandrasekharan et al., 1988)

Mass balances of tracer are very useful because, as discussed above, they can reveal if some of the tracer (and recharge) have moved by a different pathway, for example along fissures or root channels. Repetitive sampling of the same profile is also useful in this context. For example, Smith et al. (1970) showed by repetitive profiling in the English Chalk that some tritum was lost from the unsaturated zone, and implied that 15% of recharge bypassed the matrix where piston flow was occurring.

11.6.3 Throughput method (chloride)

If a tracer does not have a signature that dates the profile, recharge can only be estimated if the concentration of the tracer is affected by evaporation processes that reduce the precipitation to recharge. Conservative, ie non-evaporated, tracers are of course concentrated in recharge as evaporation proceeds, so that the flux of tracer input at the surface equals the flux of tracer reaching the water table. As the age of the tracer reaching the water table is unknown, it is necessary to assume a steady input of tracer at the surface. Only environmental tracers have been used, as a long enough input is needed to reach steady state; chloride is the most commonly used ion.

Assuming no input of tracer from minerals and that water and tracer are transported at the same rate, the flux balance of tracer between surface and water table is:

$$r\ T_r = p\ T_p + f_d \qquad 11.26$$

(Notation below eqn 11.27.)

The natural sources of chloride are from the ocean and terrestrial processes. Another major source (and sink) of chloride is agriculture, with deposition from fertilisers and animals, and removal in crops, soil erosion and animals. Agricultural fluxes of chloride can dominate the balance, and the method should be used with care in such areas.

Omitting dry deposition:

$$r = p\ T_p/T_r \qquad 11.27$$

where r : mean recharge rate (L/T)
T_r : mean tracer concentration in recharge (M/L^3)
p : mean precipitation (L/T)
T_p : mean tracer concentration in precipitation (M/L^3)
f_d : dry deposition flux (M/L^2T)

The method therefore consists of measuring precipitation, tracer concentration in precipitation, and tracer concentration profiles with depth from cored boreholes. Provided recharge and tracer inputs are in steady state and there is no secondary recharge mechanism, tracer concentration should increase with depth until a steady concentration is reached (Fig. 11.10a). This indicates that no evaporation takes place from below this depth and eqn 11.27 can be applied.

Profiles such as Fig. 11.10b-d must be interpreted with care. They indicate that one of the assumptions of steady state, no secondary pathways and no input from minerals or agriculture do not apply.

Allison & Hughes (1983) interpret profiles like Fig. 11.10b as a change in recharge rate at some time in the past, in their case due to a change in land use. Allison et al. (1985) regard another profile like Fig. 11.10b as showing a change in recharge rate some 16,000 years ago. In both cases, the recharge rates are so low as to be of little interest for groundwater resource modelling (see Chapter 1).

Sharma & Hughes (1985) use profiles like Fig. 11.10c by assuming that concentrations at the water table can be used in eqn 11.27. Other examples of the use of chloride profiling are given by Sharma (1988) and Edmunds et al. (1988).

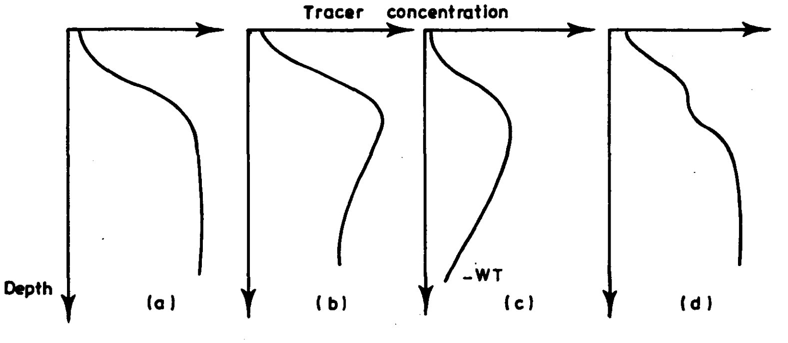


Fig. 11.10 Schematic profiles of a conservative tracer. (a) Ideal case, recharge can be estimated. (b) Probable change in recharge history. (c) Secondary pathways for recharge. (d) Addition of tracer from soil

11.6.4 Sampling methods for tracers

All methods using environmental tracers need records of their input, often over long periods. The isotopic composition of rainfall at monthly intervals for 100 selected stations throughout the world has been published since 1953 by the International Atomic Energy Authority (IAEA, 1969-1986). Local data are usually needed for any study, but the published data can be used to extrapolate short records.

Both signature and throughput methods of using tracers need profiles of pore water concentration with depth. A cored borehole is therefore needed for each profile, unless an applied radiotracer is to be detected in situ from an adjacent borehole.

Sampling for tritium is particularly difficult because samples can easily be contaminated, especially at low moisture contents when there is very little tritium present. Details of isotope sampling techniques are given by IAEA (1983).

Chloride sampling is more robust because concentrations increase as recharge rates fall. The British Geological Survey have reported on a number of low cost methods of obtaining chloride samples. These are detailed in Section 20.3.

More details of field techniques are given by Athavale & Rangarajan (1988 and Chapter 18).

11.7 Variability of recharge across catchments

Recharge varies across catchments because the controlling factors vary, both in their nature and size. These factors include:

- precipitation and other water supplies
- geology and soil
- vegetation and land use
- topography and landform
- groundwater condition.

Because recharge is a non-linear process, it is not possible to use average values of each controlling factor to derive an average recharge. Recharge should be estimated separately for each homogeneous zone; the spatially varying values are of course essential for groundwater modelling studies.

In general, the more detailed the subdivision into zones, the more accurate will be the recharge values, but the more expensive and time consuming will be their estimation. The amount of data available (maps, climate stations, Landsat imagery, etc) will partly determine the detail that can sensibly be achieved. Sensitivity analyses will help decide how important are variations in the controlling factors. Even where data on precipitation and climate are scarce, zoning on the basis of existing data will still be valuable. It will enhance understanding of processes in the catchment, and allow subjective adjustments to recharge estimates.

Table 11.4 suggests a set of factors that should be used to zone a catchment for recharge calculations. Initially, each factor should be mapped on a separate overlay, then the overlays combined to produce a master map of homogeneous zones. Recharge is then estimated for each zone by one of the methods discussed in Sections 11.1-11.6 above.

Some authors have observed important changes in recharge controlling factors over very short distances (<1 to 100 m) in apparently homogeneous terrain (eg Berndtsson & Larson, 1987; Nielsen et al., 1973). Others have reported significantly varying recharge over similarly small distances (eg Sharma & Hughes, 1985; Kitching et al., 1977). The importance of these variations has not yet been established, and there is certainly no practical way to take account of them for engineering studies. Three to ten repeat samples within each zone (as defined by the factors in Table 11.4) would certainly be worthwhile if they can be afforded, but first priority should go to obtaining one value from each zone.

Table 11.4 Factors for classifying recharge zones

Factor	Example values or classification scheme
Precipitation type	monsoon/thunderstorm/winter/summer/etc.
Precipitation amount	20% increase in pptn between zones.
Irrigation type	sprinkler/furrow/paddy/flood/canal/etc.
Irrigation amount	as precipitation.
Evaporation potential	20% increase in Penman potential evapotranspiration between zones. Solar or net radiation can be used in arid and semi-arid areas.
General lithology	alluvium/chalk/limestone/sand/etc.
Soil classification	Important factors are infiltration capacity, moisture storage and depth. Natural vegetation or drainage density may be adequate substitutes if no soil data available.
Land cover	grass/arable (crop type)/natural forest/ plantation/phreatophyte/urban/etc.
Landform	floodplain/rolling hills/plateau/etc.
Depth to groundwater or capillary fringe	> root depth/ within reach of plants/ within soil zone/ direct evaporation possible.

Notes
This classification is not intended to be exhaustive. On the other hand, it will often be unnecessary to use all classifying factors.

Data sources:
- topographic maps
- geological and soils maps
- Landsat and other satellite imagery
- aerial photography
- climate maps and meteorological stations
- precipitation records
- irrigation scheme maps and records
- field visits!

11.8 Localised recharge

The introduction to this chapter pointed out that there is a category of recharge intermediate between direct (at the spot where the precipitation falls) and indirect (along main river channels). This intermediate category is called localised, implying some horizontal movement of water before recharging groundwater. This movement is on a scale too detailed to map for engineering studies, and therefore causes great difficulty in estimating recharge. Unfortunately this type of recharge is often the largest in arid and semi-arid areas.

A number of situations where localised recharge will occur are identified in the following sections. Suggestions are made about the most appropriate methods for estimating recharge in each.

11.8.1 Hardrock terrain

In weathered, bare hardrock or limestone terrain, recharge is into distinct fissures; Chapters 7 and 8 give some examples. There are few satisfactory ways of estimating the localised recharge component in such terrain. Tracers, Darcy's law below the water table, and representative basins offer the best possibilities. Otherwise, consider using methods of estimating net recharge to a region, for example measuring catchment discharge; these methods are discussed in Chapter 16.

11.8.2 Topographical depressions

Topographical depressions can range in size from centimetres to enclosed catchments of several square kilometres. Studies in a wide range of terrains has shown that recharge is focussed in such depressions, where they exist (eg, Freeze & Banner, 1970; Kafri & Ben Asher, 1978; Rehm et al., 1982; Harhash, 1980). This is true even for highly permeable materials.

At the smallest and intermediate sizes, there is no alternative but to ignore the depressions.

The largest sizes are typified by Qatar, where 7000 km^2 or 61% of the total land area drains to 850 depressions; individual catchment areas range from 0.25 to 45 km^2. Estimating recharge in such cases is similar to estimating it from minor wadis (Section 11.8.3), but is simpler. The first activity must be to estimate the amount of runoff.

In Qatar, recharge was estimated by water balance methods for the period of detailed observation (Harhash, 1980; Lloyd et al., 1987). Rainfall data and post-storm surveys of individual depressions were used to estimate rainfall-runoff characteristics; 15-25% of rainfalls over 10mm/day became runoff. Water level measurements showed how long depressions took to empty by evaporation (estimated by the Penman open water formula) and infiltration. Allowances were made for soil moisture storage in depressions, which would later be evaporated. Once a relationship between rainfall and recharge had been developed for the period of observation, it was used to extrapolate backwards in time when only precipitation data were available.

Other techniques which have been successfully used in such terrain are Darcy's law, tracers and groundwater response (Chapter 16).

11.8.3 Minor wadis

In arid and semi-arid areas there are many small channels which carry ephemeral flows only in response to significant storms. These flows will often not reach the main channel, where flow gauging stations are usually located, but will infiltrate into the channel alluvium. These channels may drain small bare rock outcrops, in which case they will hardly be visible, or may drain catchments of several thousand km² in areas where little flow gauging is carried out. The particular case of "mountain front" recharge is considered in Section 11.8.4.

The most common approach to estimating recharge in these cases in practice is a mixture of water balance and empirical formula. Rainfall-runoff relationships are estimated or transposed from other areas, and the proportions of runoff going to each destination (evaporation, soil moisture storage, irrigation, etc) is measured, estimated or guessed; the residual is recharge.

The most satisfactory approach would be to collect field data from a number of locations, using the representative catchment approach discussed in Section 11.3. This could be supplemented by environmental isotope and groundwater response data.

11.8.4 Mountain front recharge

Many alluvial aquifers can be described as mountain front systems, including alluvial fans, piedmont plains and subsidence basins (Section 3.2). The major sources of recharge to these systems are often along the mountain boundary, consisting of two components:

(i) subsurface inflow from the mountain mass to the basin infill,

(ii) infiltration of runoff in defined channels, minor unmapped channels, and from mountain slopes.

These are illustrated in Fig. 3.2.

The two components of mountain front recharge are difficult to separate in practice. The runoff component can be estimated by water balance methods, as discussed in Section 11.8.3, but the subsurface flow is difficult to isolate. The best solution is probably to estimate the total by a Darcy throughflow calculation within the main aquifer. The section over which flow is estimated should be well within the aquifer, away from the mountain boundary, for two reasons. All of the mountain front recharge will occur upslope of the section as all the wadis flows will have infiltrated. Secondly, it is probable that more data on groundwater gradients, hydraulic conductivities and aquifer cross-section will be available in the main body of the aquifer, so giving greater accuracy and confidence in the results.

This section is based upon an excellent review of mountain front recharge by Wilson in WRRC, 1980.

12 RECHARGE FROM RIVERS

12.1 Introduction

Recharge from rivers is probably the most difficult type of natural recharge to estimate. There is more variability in flow than in precipitation, as much variability of aquifer controls as for irrigation, and greater difficulties of measurement than either. This chapter attempts to catalogue approaches to the problem, but cannot cover every possibility. The most important part of any technique is to understand the hydrogeological context before making any calculations.

12.1.1 River types

Rivers in arid and semi-arid areas can be classified in several ways, for example by flow characteristics:

(i) Perennial, that is flowing all year. This implies a river source in a higher rainfall catchment, or a river fed by groundwater.

(ii) Seasonal, that is flowing for part of the year.

(iii) Ephemeral rivers flow only in response to storms and are the classical type found in arid areas.

An alternative grouping is by their connection to water table. This will affect the ability of the aquifer to accept water and hence may control recharge rates. Three broad types can be defined:

(i) Remote from (high above) the regional water table. In general, perennial and seasonal rivers must be perched, that is the river runs in a low conductivity material, while ephemeral rivers need not be perched.

(ii) Connected to groundwater.

(iii) Above the water table, but close enough for the water table to rise to the river in response to recharge. For the purposes of this chapter, this intermediate category has been included with (ii) connected rivers.

Different methods of recharge estimation apply to perennial/seasonal rivers on one hand and ephemeral rivers on the other. Remote rivers require differences of approach from those connected to the water table. Rodier (1984) discusses the nature of arid and semi-arid rivers in various climates and parts of the world.

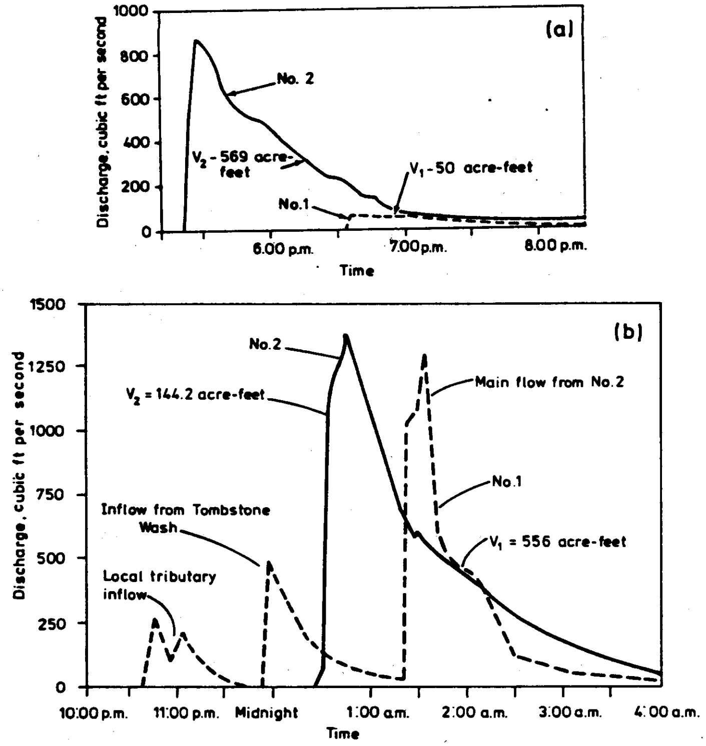


Fig. 12.1 The effect of antecedent conditions on transmission losses. (a) High losses on 10th August 1959 (b) Low losses on 3-4th August 1959. No. 2 is the upstream gauging station, No. 1 is downstream (Keppel & Renard, 1959)

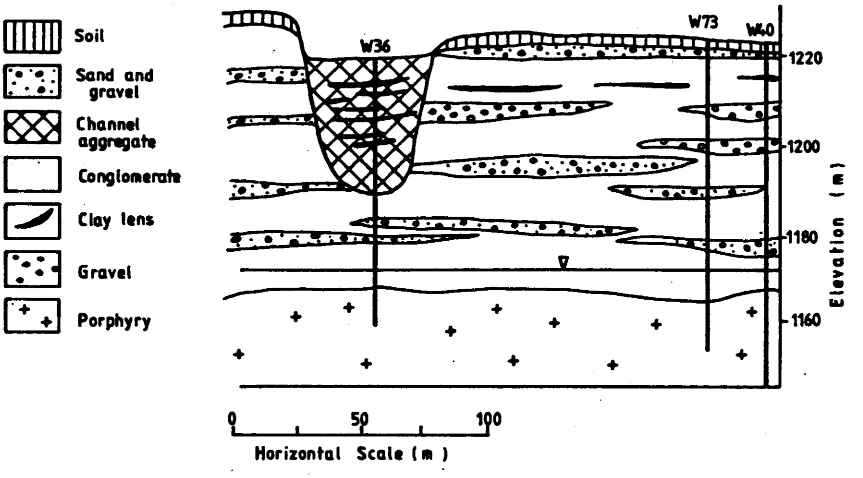


Fig. 12.2 Cross-section through Walnut Gulch, Arizona, showing typical complexity of alluvial geology (Wallace & Renard, 1967)

12.1.2 Controls on river recharge

The amount of recharge is controlled by the river flow, the river bed, and the aquifer. Controlling factors may include:

> river flow: flow rate, depth, flow volume, peak flow, velocity, duration of flow, frequency of flows, temperature (affects hydraulic conductivity), silt content;
>
> river bed: antecedent conditions, width, hydraulic conductivity;
>
> aquifer: boundaries, depth to water table, hydraulic conductivity (especially variations in conductivity), moisture retention in unsaturated zone.

As an example of such controls, Fig. 12.1 shows the effect of antecedent conditions on transmissions losses in a perched ephemeral river. The first flood (Fig. 12.1a) occurred when the channel was dry and almost all flow is lost. In the second event (Fig. 12.1b) there are tributary inflows between the two gauging stations before the main flood event arrives at the downstream end. These wet the channel bed and the transmission losses are much lower.

Fig. 12.2 shows a cross-section through the Walnut Gulch channel in Arizona. The geology between river water and main water table is clearly a major control on recharge from remote rivers on alluvial materials. The following example emphasises this point.

Fig. 12.3 shows the importance of low conductivity layers in the unsaturated zone beneath a remote river. The water table responds directly to flood events (Fig. 12.3b) but water is also stored in the unsaturated zone above two low conductivity layers; Fig. 12.3b shows the variation in the elevation of the top of one perched mound, while Fig. 12.3c shows profiles of moisture content with time from ground to water table. The perched mounds do not finally disappear until several months after the flood event.

Not all the water that leaves a river to move downwards becomes recharge. The distinction between transmission losses, deep percolation and recharge is useful. Transmission losses are the river flows that don't arrive at the downstream end of a river, deep percolation is the water that enters the aquifer, and recharge is that which crosses the water table. Transmission losses are reduced by bank storage and evaporation from the surface and temporary pools before becoming deep percolation. This in turn can be reduced by evaporation, perched water tables and underflow before becoming recharge.

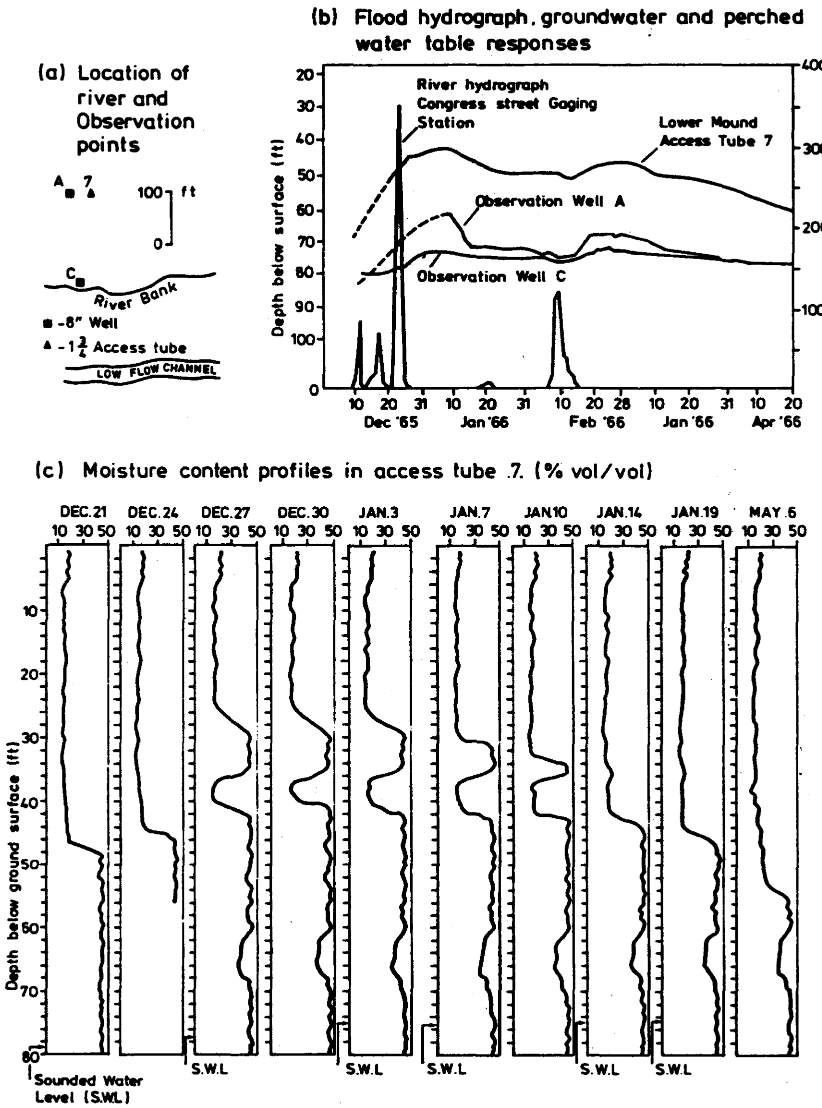


Fig. 12.3 Effect of low conductivity layers on river recharge, with perched water tables forming at two levels (Wilson & De Cook, 1968)

12.1.3 General procedure for estimating river recharge

With all these factors it is clear that recharge cannot be estimated by considering the river alone. A general procedure would be to:

(i) consider how much water can be accepted by the aquifer, that is how far the water table can rise and how fast water can flow away from a recharge mound;

(ii) estimate the transmission capacity of the unsaturated zone, if there is one. Knowledge of low conductivity layers (which will cause perching) will be essential;

(iii) finally consider the river flow and river bed processes, estimating possible deep percolation and all the other items of a water balance such as evaporation from the bed.

Recharge from remote rivers is mainly controlled by (ii) and (iii), the river and unsaturated zone conditions. Rivers connected to groundwater are primarily controlled by (i), the groundwater conditions.

12.1.4 Numerical modelling of groundwater systems with connected rivers

The ideas of potential and actual recharge discussed in Sections 11.1.2 (for precipitation) and 14.1.1 (irrigation) are also relevant to river recharge, especially for rivers which are connected to groundwater (Rushton, 1987). Groundwater conditions partly control recharge from connected rivers. As groundwater conditions change, for example with pumping from new boreholes, recharge will change. Estimates of historical average recharges would not be valid under such changed conditions.

Estimates of historical, ie actual, recharge are needed for calibrating groundwater models. On the other hand, for predictive modelling it is essential to know what the recharge would be if groundwater conditions changed; this quantity is clearly different from actual historical recharge, but may not be the same as potential recharge.

The best way to proceed is:

(i) estimate historical recharges by one of the methods discussed in this chapter;

(ii) calibrate the groundwater model treating the river as a specified flux boundary, that is using these recharges as inflows to the model;

(iii) once the aquifer section of the model is calibrated, alter the model to treat the river as a leaky boundary, that is where the model calculates recharge from both river and groundwater levels. Rushton and Tomlinson (1979) discuss possible relationships to use. The model must keep an account of the water left to flow downstream in the river to ensure that recharge does not exceed river flow;

(iv) calibrate the new river part of the model to reproduce the historical recharges;

(v) use the model in predictive mode. It will now estimate recharge for future groundwater conditions.

12.1.5 River recharge estimation methods

Direct measurement of river recharge is not possible. A number of empirical methods have been devised and are discussed in Section 12.2. Included here are a number of methods based on groundwater response. Water balance methods are also possible for all three items, transmission loss, deep percolation and recharge (Section 12.3). Darcian approaches, that is based on the groundwater flow equation, are possible in theory but are difficult to use in practice with important simplifications (Section 12.4). Tracer techniques have very limited use for quantifying river recharge (Section 12.5). Models of river hydraulics and of the complete hydrological cycle in catchments have been used for river recharge estimation and these are discussed in Section 12.6. Fig. 12.4 outlines the choices for the various river conditions, and Table 12.1 summarises the methods outlined in this chapter.

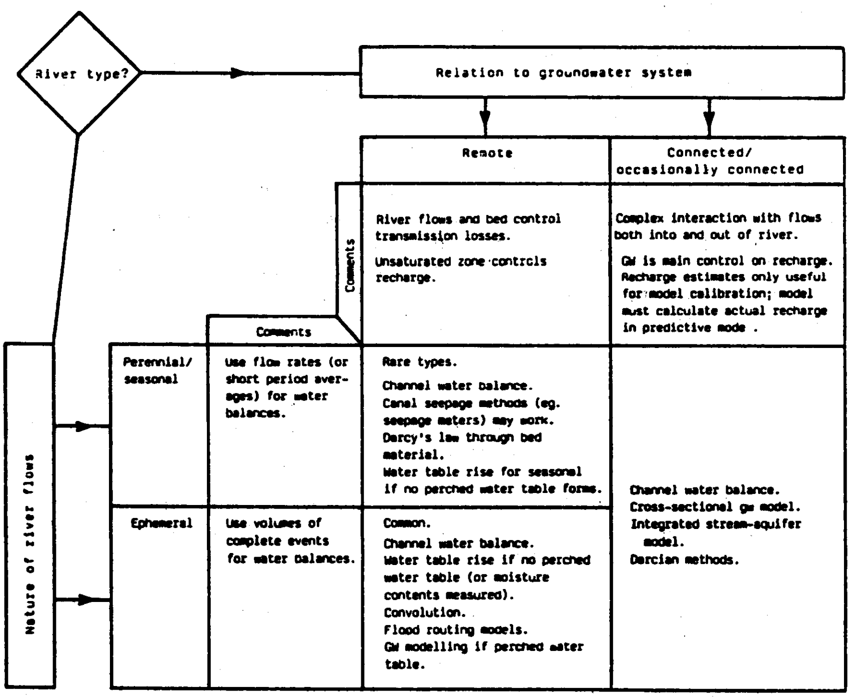


Fig. 12.4 Relation of type of river to recharge estimation methods. See also Table 12.1

12.2 Empirical methods

In an empirical method, recharge is related to one or more of the controlling variables, for example river flow rate. Some examples are shown in Fig. 12.5.

Table 12.1 Comparison of methods of estimating river recharge

This table is not a substitute for the longer discussion of the methods given in the main text of this chapter.

Method	Empirical formulae	Channel water balance	Darcian approaches: numerical modelling of river-groundwater system	Tracer profiling methods	Surface water modelling (FR - flood routing CM - catchment modelling)
Section	12.2	12.3	12.4	12.5	12.6
Applicability	Transmission losses in remote rivers	Transmission losses in all rivers	Recharge from all rivers	Little value for quantitative work	FR - transmission losses, remote ephemeral rivers CM - all rivers
Accuracy	Moderate	Moderate - controlled by accuracy of measurements. More accurate for higher ratio of recharge to flow	Theoretically good, in practice very variable. Must be calibrated		FR - moderate to good CM - poor to moderate
Data requirements	Independant estimates of losses or recharge to calibrate formula. Continuous records of river flow or gw levels as appropriate	Coninuous records of up and down stream flows and of all other in and out flows along reach	Conceptual model of system. Aquifer properties (K, S, K-ψ-θ relations). Calibration data (eg unsaturated and gw heads). River flows or heads		Continuous flow records at several sections for calibration FR - input flows for period of interest CM - meteorological data
Ease of use	Simple, once formula established; convolution methods more difficult to establish	Simple arithmetic	Complex: generally considered as research tools. Preparation of computer code, aquifer properties, and use of models can be difficult and time consuming		Moderate to difficult
Type of estimate	Total over river reach for timestep of analysis (event or season)	Total for timestep used (day, month, season, etc)	Instantaneous point or cross-section value		FR - transmission losses for every reach CM - net recharge every reach for each time step
Costs	Low if gauging station etc already operating	Low if gauging stations already exist, high ($50 000+ otherwise)	High for specialist modelling skills and field instrumentation.		Cheap if data available, otherwise expensive
Time required	Short	Rapid if data available, otherwise minimum of a year	Moderate to long, 6-24 manmonths per section for modelling		Moderate (3-12 manmonths) if data already available long otherwise

There are several general comments about the usefulness of these methods. Firstly they are likely to work best for transmission losses in perched rivers, for which there is less influence of the unsaturated zone and aquifer. Good estimates of recharge are needed to develop such relationships; once data collection networks have been installed, is it not better to continue to use them than to rely on a derived relationship? Finally, the relationships can only be transferred to other rivers which have the same conditions (see list above in Section 12.1.2).

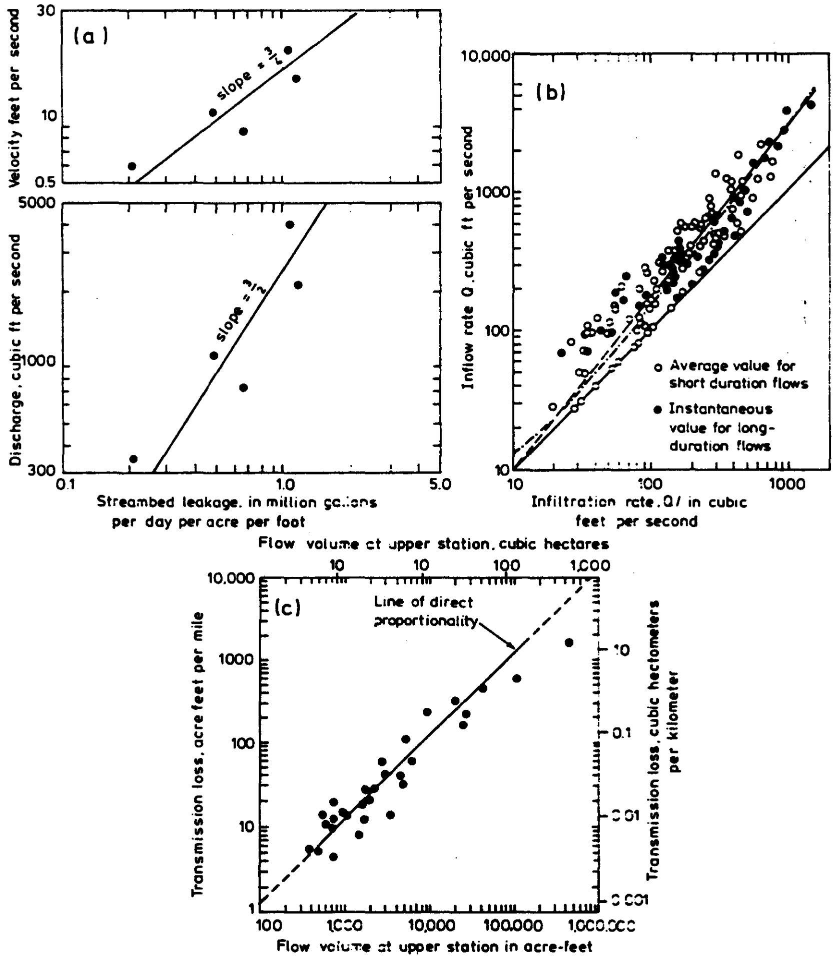


Fig. 12.5 Examples of empirical relationships for transmission losses from remote rivers. (a) Norris, 1970 (b) Burkham, 1970 (c) Jordan, 1977

12.2.1 Transmission losses in ephemeral rivers

A detailed study of perched ephemeral streams in Kansas was carried out by Jordan (1977), who studied 53 events spread over pairs of gauging stations on 14 rivers. He worked with the volumes of complete events and found that transmission losses correlated well with flow volume at the upstream station (Fig. 12.5c). He also showed that losses along the channel varied exponentially

$$V_x = V_0 R^x \qquad 12.1$$

where V_x : flow volume at distance x downstream (L^3)
V_0 : flow volume at upstream end of reach (L^3)
R : ratio of flow volumes 1 length unit apart

Table 12.2 Summary of transmission loss data for alluvial channels in Tucson Basin, Arizona. (Based on WRRC, 1980, Table 5.4)

Stream	Infiltration Rate (m/d)	Flow Loss ($10^6 m^3/km/y$)	Comments
Santa Cruz, Continental Tucson	0.91-2.0		Calc. from streamflow losses
Santa Cruz, Continental Tucson		0.25 (45% of inflow)	Calc. from streamflow losses
Santa Cruz, Tucson-Cortero	1.9-2.2		Calc. from streamflow losses
Rillito	0.3-1.1		Snowmelt runoff calc. from seepage runs
Rillito	2.4	3.9	Snowmelt runoff calc. from gw level changes
Rillito		0.63 (56% of inflow)	Calc. from streamflow losses
Rillito Creek	1.3-3.3		Calc. from streamflow losses
Pantano Wash		0.18 (72% of inflow)	Calc. from streamflow losses
Pantano Wash	1.0-1.2		Calc. from streamflow losses
Tanque Verde Creek		0.33 (45% of inflow)	Calc. from streamflow losses
Sonoita Creek	2.4-1.8		Low on flood plain High in stream channel
Queen Creek	0.35-5.2 (ave. >1.2)		Seepage measurements winter flow event
Salt River	0.46-0.76		Calc. from streamflow losses
Rincon Creek		0.34 (92% of inflow)	Calc. from streamflow losses

Other examples of empirical relationships for transmission losses are given by the papers from which Figs 12.1, 12.2 and 12.5a+b were taken, and by Lane et al. (1971). WRRC (1980) reviewed a large number of American studies of transmission losses in remote ephemeral rivers. Table 12.2 summarises observed loss rates. They suggested some general rules for transmission losses which are summarised in Chapter 3.

12.2.2 Groundwater response in ephemeral rivers

There are a number of empirical techniques based upon analysing groundwater response. They are typified by the convolution approach which is outlined below, and include a spectral analysis method (Gelhar et al., 1979) and a linear transformation of the annual distribution of rainfall (Flug et al., 1980). The latter two require more simplifying assumptions but are more mathematically complex than the convolution approach and so are not described here.

All of the groundwater response methods require a good independant estimate of recharge for at least one event. Without this, the parameters of the method cannot be found. Therefore any results for other events will be less accurate than the one independant estimate.

The convolution method has been described by Moench & Kisiel (1970), Hall & Moench (1972), Besbes et al. (1978) and others. This account draws heavily on Besbes et al. who give a thorough comparative study of several methods in a semi-arid area of Tunisia.

The groundwater flow equation is commonly considered linear, that is the principle of superposition applies. If I(t) is the water table response to a unit input of recharge in one time unit, 2 I(t) is the response to 2 units of recharge in one timestep, and I(t) + I(t-1) is the response to two successive unit inputs. In discrete form:

$$G(i) = \sum_{j=1}^{i} R(j)\, I(i-j+1) \qquad 12.2a$$

which is equivalent to:

$$G(i) = \sum_{j=1}^{i} R(i-j+1)\, I(j) \qquad 12.2b$$

(Notation below eqn 12.3.)

Besbes et al. describe two methods to find the unit response I(t). In one, a groundwater flow model for the aquifer is calibrated, then stimulated with a unit recharge and its response observed. (This does seem a little inverted because they must of course have included a river recharge estimate to calibrate the model!) The other method is to observe the water table response to a single, large, isolated recharge event which is treated as of unit duration. In this case

$$G(t) = h(t) - d(t) \qquad 12.3a$$

and

$$I(t) = G(t)/V \qquad 12.3b$$

where

G(t) : the groundwater level rise at time t (L)
R(j) : the recharge volume in time interval j (<=i) (L^3)
h(t) : the groundwater level at time t (L)
d(t) : the groundwater level that would have occurred in the absence of recharge, from the recession curve (L)
I(t) : the ordinate of the unit response at time t (L^{-2})
V : the volume of recharge (L^3)

In practice, Besbes et al. observed the initial and peak response and constructed the remainder using a model. Examples are shown in Fig. 12.6. The volume of this recharge event was initially estimated by a water table rise calculation using a guessed specific yield, then refined by calibrating the model. Other authors have used transmission losses to estimate recharge volume.

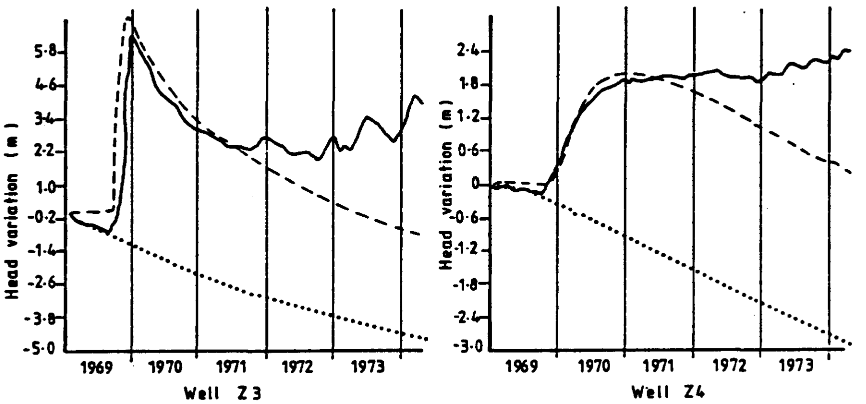


Fig. 12.6 Groundwater responses to a single large recharge event, Kairowan, Tunisia (Besbes et al., 1978). Solid lines: observed heads, Dotted lines: heads computed by groundwater model assuming no recharge, Dashed lines: heads computed with estimated recharge

Once the impulse response is known, eqn. 12.2 can be used to deconvolve the observed fluctuations in any observation well to estimate recharge. The system of equations is in fact over-determined, and a constrained least squares method is recommended.

12.2.3 Rivers in contact with the water table

Recharge from rivers that contact the water table is effected by conditions below that water table. In these cases, the empirical methods outlined above rarely work. Fig. 12.7 shows data for a perennial river crossing a major alluvial aquifer in N. Africa. There is no clear relation between transmission loss and river flow because of the effect of the groundwater system. No general relationship can be devised for these rivers and recharge for each time period must be explicitly calculated.

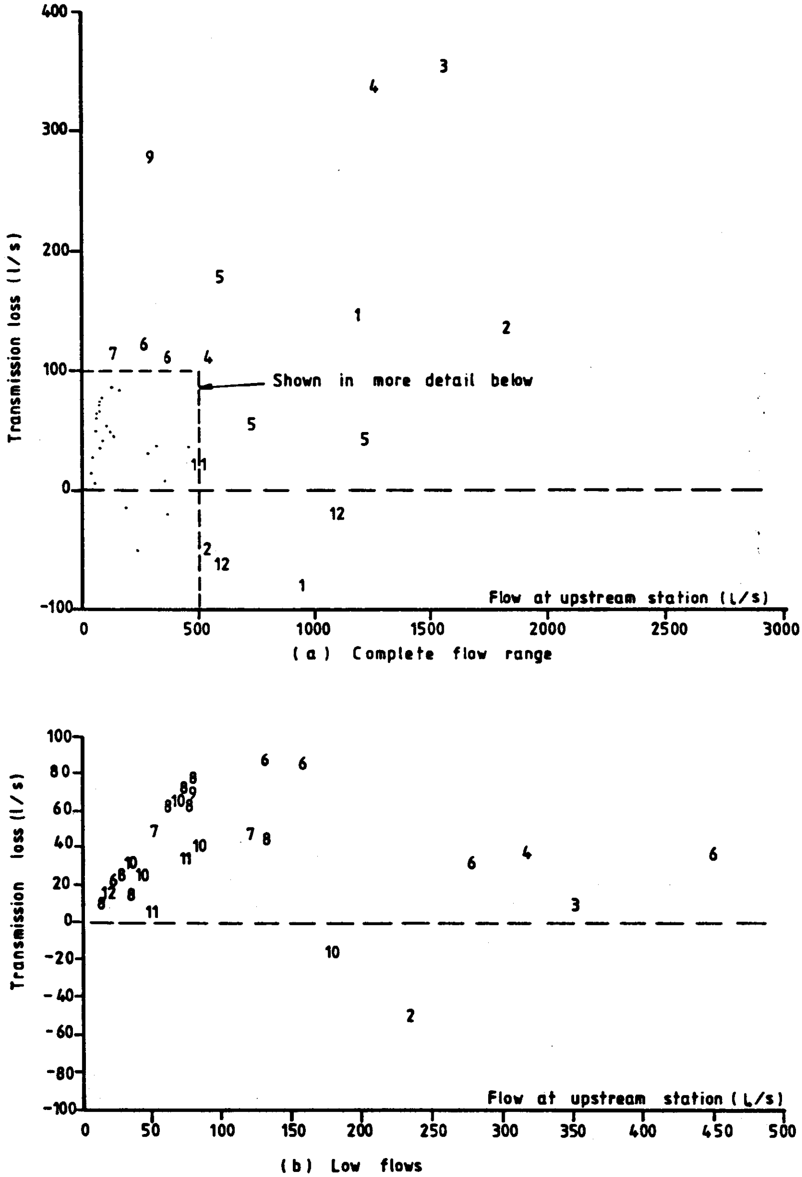


Fig. 12.7 Transmission loss -vs- flow for a river in contact with the water table showing little correlation. Each point is one month, identified by its month number (eg 1=January)

12.3 Water balance methods

12.3.1 Channel water balance

A water balance of the flows along a river reach is the most straightforward way to estimate transmission losses. By careful consideration of processes, this may be extended to deep percolation or recharge. Using river flow rates,

$$R = Q_{up} - Q_{down} + \Sigma Q_{in} - \Sigma Q_{out} - E_a - \delta S/\delta t \qquad 12.4a$$

or in volume terms,

$$R\ \delta t = V_{up} - V_{down} + \Sigma V_{in} - \Sigma V_{out} - E_a\ \delta t - \delta S \qquad 12.4b$$

where Q : flow rate (L^3/T), V : flow volume (L^3)
Q_{up}, V_{up} : the flow at the upstream end of the reach
Q_{down}, V_{down} : the flow at the downstream end of the reach
Q_{in}, V_{in} : inflows from tributaries, urban effluents and irrigation returns
Q_{out}, V_{out} : outflows for water supply, irrigation
E : evaporation from water surface or stream bed (L^3/T)
δS : the change in channel or unsaturated zone storage (L^3)

Total flow volumes (eqn 12.4b) should be used in preference to instantaneous flow rates for ephemeral rivers, as lag between the two gauging stations will be difficult to estimate accurately.

Any time period can be used with eqn 12.4a for seasonal and perennial rivers, but there are advantages in using a short period. More data points are generated, and conditions are relatively constant during each period. This may make the process controlling recharge more obvious. More importantly, the higher errors associated with the higher flows will not dominate the whole period as would occur if averages were taken over long periods, including low flow periods. Travel time of flood waves between gauging stations should be allowed for if short time peiods are used.

<u>Evaporation</u> from the river surface can be estimated by Penman's open water method (FAO, 1977). Evaporation from ephemeral river beds becomes increasingly important as the frequency of flooding decreases. Sorey and Matlock (1969) carried out some small scale lysimeter experiments on evaporation from streambed sands. Over 19 days, total evaporation ranged from 5mm in a very coarse sand to 26mm in a medium sand.

<u>Advantages and disadvantages.</u> Channel water balances are probably the most accurate way to estimate recharge, and often provide the data on which other methods are based. They are essential when modelling rivers which may receive as well as donate water. The data collected may have many other uses, including rainfall-runoff modelling for transposition to ungauged catchments.

The disadvantages include cost and accuracy. Setting up, maintaining and running river gauging stations is expensive. They must be able to measure large floods accurately. In arid and semi-arid areas where access may be difficult, floods are infrequent and river beds shift, it may be too expensive. Rodier & Roche (1978) discuss the difficulties of measuring river flows in arid and semi-arid areas.

The method is prone to inaccuracy as recharge is calculated as the difference between large numbers. Measurement errors on high river flows are commonly ±25%, and for flash floods in arid and semi-arid areas can easily be +100%, -50%. Thus the possible errors in recharge estimates can be ±100%.

12.3.2 Water table rise

An alternative to a surface water balance for transmission loss is a groundwater balance for river recharge. Observation wells perpendicular to the axis of the river will show the profile of a recharge mound. Under ideal conditions the volume of water in the mound is the amount of recharge. Several points should be noted:

(i) For perennial or seasonal rivers, the groundwater flows away from the mound over the recharge season may be significant in relation to the change in storage.

(ii) Groundwater flows away from connected rivers may be large compared to any change in storage.

(iii) Storage in the unsaturated zone is common below ephemeral rivers on alluvial aquifers. Wilson and De Cook (1968) found that 33% of recharge showed up immediately as a water table rise; the remainder took several months to arrive (Fig. 12.3). Measurements of moisture content are essential in such cases, for example, by neutron probe.

(iv) The water table rise should be measured relative to the recession that would have occurred if there had been no recharge.

(v) Specific yield is a difficult parameter to estimate. It can show hysteresis, and have unexpected changes due to aquifer heterogeneity - see Section 16.2.

12.4 Darcian approaches

There are three ways to use Darcy's law in estimating recharge from rivers, as follows:

(i) using infiltration equations and flow nets, that is desk studies of homogeneous and isotropic aquifers with assumed boundary conditions,

(ii) gathering field data on aquifer properties, moisture content and pressure in the unsaturated zone, and piezometric heads in the saturated zone, and using these data in the equations of flow to analyse real events,

(iii) using numerical models.

12.4.1 Infiltration equations and flow nets

The Darcian approach is typified by the infiltration technique (see Fig. 12.8). An ephemeral stream floods a dry river bed, separated from the water table by an unsaturated zone in which there is no flow initially. Infiltration occurs and a saturated zone extends downwards. The edge of the saturated zone is assumed to be sharp and to mark the division between flowing and stationary water; it is commonly called the wetting front. The Green & Ampt solution for the infiltration rate is

$$I(t) = K\ [H(t) + L(t) - h_c)\ /\ L(t) \qquad 12.5a$$

$$L(t) = V(t)\ /\ (\theta - \theta_i) \qquad 12.5b$$

where
- $I(t)$: infiltration rate at time t (LT^{-1})
- K : resaturated hydraulic conductivity, taken as half of the fully saturated value (LT^{-1})(Bouwer 1978)
- $H(t)$: depth of water ponded on the surface (L)
- $L(t)$: depth of the wetting front below the river bed (L)
- h_c : air entry head, representing the height of the capillary fringe (L)
- $V(t)$: depth infiltrated up to time t (L)
- θ : resaturated moisture content
- θ_i : initial moisture content

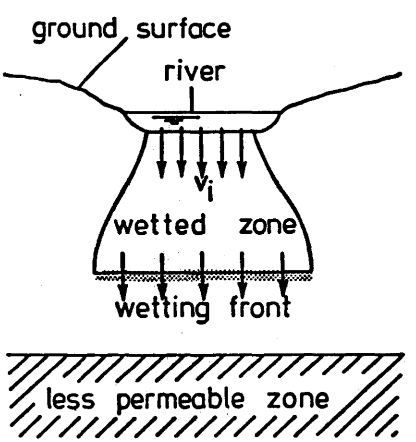


Fig. 12.8 Idealised infiltration from a perched river into a homogeneous material

The infiltration approach assumes no prefential pathways and no lateral flow, so would not be valid in the case of Figs.

12.2 and 12.3 where low conductivity layers will cause perching and lateral flow. The equation ceases to be valid once the wetting front has reached the water table, when horizontal flow occurs. Freyberg (1983) has looked at the effects of a varying wetted perimeter of the river, and shown that they can be important. Unfortunately it is difficult to generalise his results.

A flow net can be constructed to represent the flow system away from a river or canal on a cross-sectional diagram. Flow nets are only easily drawn for steady state flows. They can seriously underestimate seepage losses during times of wetting when much higher flows are going into aquifer storage.

Contours of equal head (equipotential lines) are drawn with equal head increments between them. They are intersected at right angles by flow lines so that each section of the grid is a curvilinear square (Freeze & Cherry, 1979).

To attempt this construction it is necessary to approximate the position of the phreatic surface which denotes the upper boundary of the flow net as a flow line. The flow within each flow tube is calculated using Darcy's Law and the scaled graphical construction.

Variations in permeability within the section result in a refraction of the flow lines at the permeability interface, and the refraction angle can be calculated using the tangent law (Freeze & Cherry, 1979). This allows flow nets to be used within a heterogeneous section, but the construction tends to be very approximate and should be used with great care.

Cedergren (1977) is a recommended text, containing further information with examples on the construction and use of flow nets.

12.4.2 Field data-based methods

Field data on moisture content and pressure in the unsaturated zone beneath river beds have been collected for research projects on recharge estimation (for example, Fig. 12.3). The author knows of no instances where such data have been collected for resources studies, presumably because of the expense and difficulty of collecting enough data to describe a transient process with three dimensional variability.

Head and hydraulic conductivity data from connected rivers could be used to estimate flow away from the river; velocity measurements using tracers could also be used. These are discussed in more detail in Chapter 14 because they have more relevance to irrigation recharge. The difficulty in using them for rivers is the transient, three dimensional nature of the flow pattern.

12.4.3 Numerical modelling techniques for river recharge

Dillon (1982) gives a thorough review of stream-aquifer models, quoting over 100 references. He classifies models into three broad types:

(i) Surface water models, including flood wave models considered in Section 12.6.1, and linear response models typified by the convolution approach discussed in Section 12.2.1;

(ii) Groundwater models, discussed below;

(iii) Integrated stream-aquifer models which can be coupled externally (eg. output from groundwater model is input to surface water model with no feedback) or coupled internally.

Groundwater flow models for river recharge estimation can be 1-dimensional vertical for remote rivers (ie. infiltration or wetting front models), or 1-dimensional horizontal for fully connected rivers (ie. Dupuit-Forchheimer assumptions) or 2-dimensional cross-section models for connected or occasionally connected rivers. Similar approaches are used for canal seepage and are discussed in Chapter 14.

Dillon (1982) lists and comments upon the assumptions commonly used to make such models tractable. He reports that practical studies and field observations have shown that the following assumptions, if used, can have significant effects on estimated recharge:

- homogeneous, isotropic and infinite aquifer,
- Dupuit-Forchheimer approximation,
- fully penetrating river,
- constant recharge rate,
- isothermal,
- initially dry soil,
- no resistance to air flow.

Most of these assumptions can be dispensed with by using a more complex computer code, except of course for the first. A better description of the aquifer requires more (expensive) field data. In practice this has meant that most of these models have been used for theoretical or parametric studies of hypothetical rivers, rather than real resource studies. Case studies are usually published in reports rather than scientific journals and so are hard to obtain. Anyone wishing to construct such a model should read Dillon (1982), and consider such models as Glass et al. (1977), Narasimhan et al. (1978), Reeder et al. (1980), Vauclin et al. (1979) and Youngs (1977).

Integrated river-aquifer models. A number of authors have written models which link river and groundwater flows and heads. They are for hydraulically connected rivers, and model river flow in some detail. Almost none have been used in groundwater resource studies, and all require a large data collection, programming and calibration effort. Dillon (1982) reviews most of the published models at the time of writing.

In general, this type of model is too sophisticated for recharge estimation. They are concerned with the immediate locality of the river and with short time steps. Water may well enter the aquifer on this scale that never recharges the regional groundwater system. Such interactions only confuse a regional groundwater modelling effort. A pragmatic policy for connected rivers is to draw the model boundary outside this zone of fluctuating interaction rather than at the river bed.

12.5 Tracer techniques for river recharge

Environmental tracers in groundwater seem to have little role in quantifying river recharge. Almost all of the groundwater in the vicinity of a recharging river in an arid or semi-arid area will have come from that river. Therefore the local groundwater will carry the same environmental tracers as the recharging water, and the two cannot be separated.

One could imagine designing an experiment to use an applied tracer to estimate river recharge. The rate of spread of tracer below and laterally away from the bed would be measured in an array of sampling points along, below and away from the channel. The authors are not aware of such a method being used in practice for rivers, probably because expense of adequate sampling wells and the need for relatively large tracer injections would not be justified by the accuracy of the results. Related methods have been used for irrigation canals, and are discussed in Section 14.5.

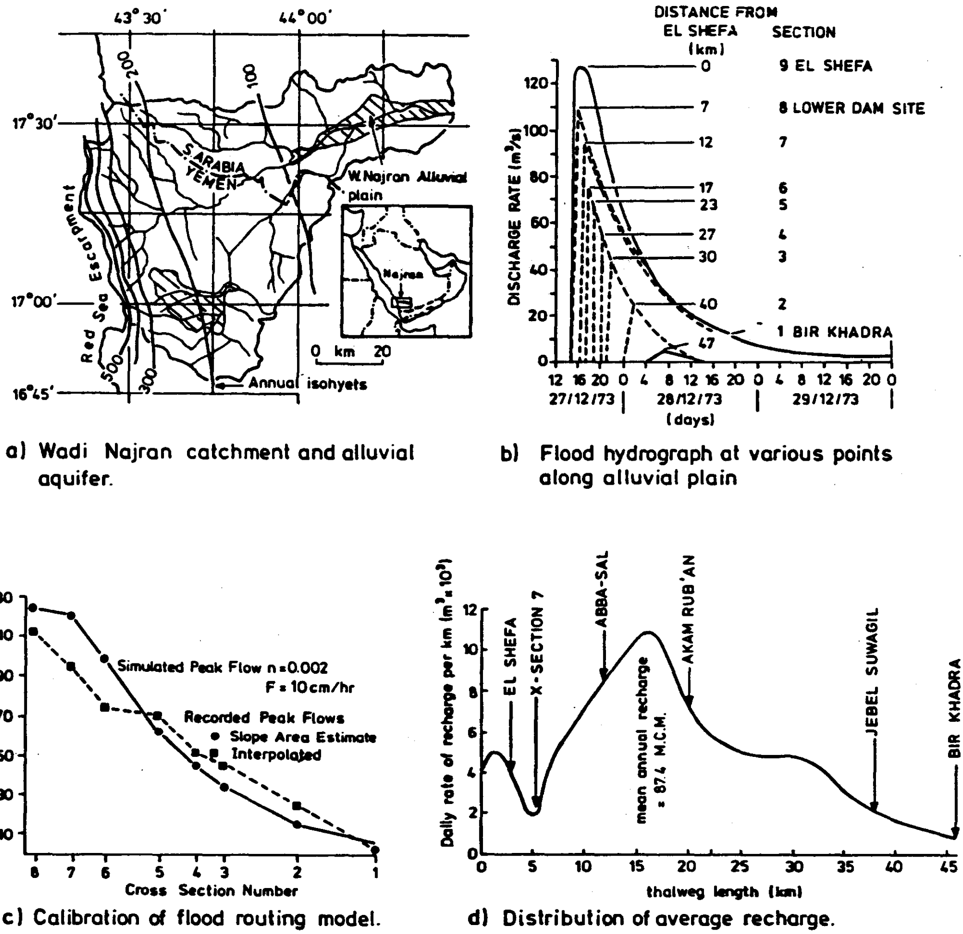
Of course isotopes and other tracers are very valuable in identifying recharge sources and in delineating zones of river recharged groundwater. Some discussion of methods is given in Chapter 20 of the Guidebook on Nuclear Techniques in Hydrology (IAEA, 1983).

12.6 Surface water modelling

There are often more data available on surface water than groundwater and so many practical recharge studies have had to rely on surface data to predict recharge. This section discusses a river flood routing method and the use of overall catchment models of rainfall and runoff. The potential errors in such methods are high as they do not usually include any aquifer or groundwater information, and are generally used when data are scarce.

12.6.1 Flood routing models

Flood routing models of river flows have been used to estimate transmission losses by Cornish (1961), Smith (1972) and Binnie & Partners (see FAO, 1981, Section 18.3.3.4). River level, or preferably flow, hydrographs are required for several stations along the river. Transmission losses and their distribution are estimated by calibrating a model of flood travel along the channel. Fig. 12.9 illustrates the method for the Wadi Najran in Saudi Arabia. In this case a single well monitored flood wave was used to calibrate transmission loss rate and Manning's n in order to match recorded flood volumes. Both the historical 7 year river flow hydrograph and two 30 year synthetic sequences were routed through the model to estimate mean annual transmission loss as 86% of inflow at the upstream end of the aquifer.



a) Wadi Najran catchment and alluvial aquifer.

b) Flood hydrograph at various points along alluvial plain

c) Calibration of flood routing model.

d) Distribution of average recharge.

Fig. 12.9 Recharge from the Wadi Najran, Saudi Arabia, estimated by a flood routing model (FAO, 1981)

12.6.2 Catchment modelling

Many practical studies have only sparse data on river flows and have been unable to carry out channel water balances. A common approach has been to develop a catchment-wide model of rainfall and runoff. In some cases this has been applied to the upstream contributing catchment to provide runoff estimates where the river crosses the aquifer boundary. The recharge process is then considered separately, perhaps using information on the frequency with which floods reach certain downstream points. In other cases the catchment model includes the aquifer and transmission losses are calibrated.

Examples include several simple models among the examples in the appendix, and models by Singh and Chowdhury (1979), Hari Krishna (1982), de Vera (1984), and Broughton and Stone (1985).

12.7 Examples of river recharge estimation

12.7.1 Introduction

The following examples are all recharge estimates for catchments where wadi infiltration is a major recharge component. A common trait of all the examples is the very approximate nature of the recharge and/or runoff estimate. This is only to be expected, considering the difficulty in both collecting the data from what tend to be very remote areas, and then assessing a mean from the sum of stochastic episodes tending to have a large standard deviation.

Cases 1 to 4 are taken from the same arid region of the Middle East and therefore have a number of features in common. The flow system within each starts with rainfall onto a hard rock upland catchment, which results in little or no direct recharge but allows a percentage of the precipitation to become runoff. The runoff flows along alluvial wadi channels, taking it off the hard rock catchment and across a lowland alluvial or sedimentary aquifer sequence. Recharge can now take place until the wadi reaches a discharge point or the water supply is exhausted.

Average annual rainfall for the upland catchment varies from 180-300mm/y and potential evaporation from 1400-2600mm/y. Rainfall in the lowland area is lower, at approximately 100mm/y, and potential evaporation is higher; this, coupled with the flat topography and the depth to the underlying water table, results in very little recharge from lowland rainfall.

All the examples are taken from project reports; references are only given where these reports are not confidential. The results of all 9 cases are summarised in Table 12.3.

12.7.2 Case 1

The hard rock upland catchment covers an area of 1650 km^2, the runoff from which flows across an alluvial coastal aquifer to the sea. It has been estimated that 8% of the runoff from the foot of the upland catchment is recharged to the wadi alluvium, the remainder recharging the coastal aquifer or being lost to the sea. The percentage of runoff resulting in recharge is dependent on the annual rainfall; in a 1 in 5 dry year 80-100% of runoff results in recharge, ranging to a 1 in 5 wet year where only 30-50% of runoff is recharged, with the rest reaching the sea.

Table 12.3 Recharge data summary for the case studies outlined in Section 12.7

Case:	1	2	3	4	Jebbel Marra
Catchment Area (km^2)	1650	405	1000	626	2470
Precipitation (mm/y)	270	180	180	186	450-650
Runoff (% Rainfall)	-	17	20	29	23-34
Recharge (% Runoff)	30-100	79	80	72-100	-
Recharge (% Rainfall)	-	13	16	21-29	-
Wadi Recharge ($10^3 m^3/km/y$)	-	138	280-430	406-493	180-470

Case:	6	Wadi Jizan	Wadi Dhamad	UAE
Catchment Area (km^2)	1800	1280	1000	4000
Precipitation (mm/y)	<40-100	480	527	357
Runoff (% Rainfall)	10	13	9-13	-
Recharge (% Runoff)	15	31	22	-
Recharge (% Rainfall)	1.5	4	2-3	14
Wadi Recharge ($10^3 m^3/km/y$)	-	500	200	-

12.7.3 Case 2

The hard rock upland catchment covers an area of 405 km^2 and receives a mean rainfall of 210mm/y. A wadi system carries the catchment runoff across an alluvial aquifer of up to 100m in thickness, into which the runoff is recharged.

Effective rainfall, equivalent to runoff for the upland catchments, is calculated for both this and the next example using the following largely empirical formulae:

$$er = Y\,(r - et_{set}) + Sf \qquad 12.6a$$

$$et_{set} = n\, et_p/365 \qquad 12.6b$$

where
- er : effective annual rainfall (L/T)
- r : annual rainfall (L/T)
- Y : factor dealing with small rainfall events, seasonal changes in transpiration, etc.
- Sf : slope factor
- et_p : annual potential evapotranspiration (L/T)
- n : number of rainfall days in the year.

The potential catchment runoff calculated by eqn 12.6 for an average year is $14.6 \times 10^6 m^3$ or 17% of the rainfall. Of this total, $3.1 \times 10^6 m^3/y$ (21%) is used consumptively in the upper catchment and the remainder recharges by two main methods, the proportions to each calculated from a simple uncorroborated model. $4.6 \times 10^6 m^3/y$ (31.5%) recharges at the foot of the hard rock and $6.9 \times 10^6 m^3/y$ (47.5%) recharges along the wadi course. With a wadi of approximately 50 km in length, this gives a recharge rate of approximately 138,000 $m^3/km/y$.

12.7.4 Case 3

This hard rock upland catchment covers an area of approximately 1000 km^2, runoff from which is carried along two main wadi courses. The wadis flow out over an alluvial plain, the alluvial sediments varying in thickness from 2-45m and overlying a low permeability basement complex. The mean annual rainfall for the upper catchment is 180mm and this results in a gross runoff for the catchment of $36.5 \times 10^6 m^3/y$ (20.3% of rainfall). Of this, $29.4 \times 10^6 m^3/y$ is recharged through the wadi beds within the study area and $7.2 \times 10^6 m^3/y$ is flood loss, flowing across the area boundary. The individual rate of recharge for the two main wadis is calculated to be 280,000$m^3/km/y$ and 430,000$m^3/km/y$.

12.7.5 Case 4

A hard rock upland catchment of 626 km^2, the runoff from which flows inland to recharge either the wadi alluvium or, beyond this, an alluvial plain.

A water balance has been attempted for a rainfall episode of 78mm ($48.9 \times 10^6 m^3$) and the balance used to estimate the recharge from an annual rainfall of 186mm ($116.7 \times 10^6 m^3$). The wadi outflow from the hard rock was estimated using Manning's equation (with a roughness coefficient of 0.035) to calculate the flow rate as follows:

$$Q = (R^{2/3} \; A \; S^{1/2})/n \qquad 12.7$$

where Q : flow rate (m^3/s)
R : hydraulic radius (m)
A : cross sectional area of flow (m^2)
S : slope of the water surface (m/m)
n : Manning's roughness coefficient.

The flows measured for the different subcatchments by this method ranged from 25-35% of the rainfall, and gave a total runoff of $14.6 \times 10^6 m^3$ for the rainfall episode. Recharge to the wadi alluvium was calculated by monitoring the flow from the aflaj (low angle wells tunnelled into the alluvium) and assuming that the outflow was proportional to the storage. Recharge registered by this method totalled $7 \times 10^6 m^3$ with another $1.5 \times 10^6 m^3$ unaccounted for in the downstream runoff and either recharged or lost as evaporation. Wadi recharge therefore ammounted to 14-17% of rainfall, which over a 40km length would give an average annual infiltration of 406,000-493,000 $m^3/km/y$.

$6.1x10^6 m^3$ was estimated to flow across the alluvial plain, where borehole monitoring and knowledge as to the specific yield of the alluvium revealed that at least $3.8x10^6 m^3$ (7% of rainfall) was recharged with the remaining $2.3x10^6 m^3$ (5%) either recharged or lost as evaporation. For a flood plain area of 150 km^2 this gives an infiltration rate of 54-93mm/y.

12.7.6 Jebbel Marra, Sudan

This study is of a 30,000 km^2 area of western Sudan, where the main aquifer system is formed by the alluvial valleys along which present day wadi flows take place. The valleys cover a total area of 2470 km^2 and are surrounded and underlain by low permeability volcanics, sandstones and Precambrian basement.

The climate is semi-arid with a mean annual rainfall of 450-650mm, most of which falls between June and September allowing the year to be divided into a recharge and a discharge season. Groundwater input during the 60 day recharge season in 1976 has been calculated by summing the increase in storage, the throughflow, and losses through abstraction and evaporation for each small alluvial aquifer over the period. The increase in storage, of $380x10^6 m^3$, was calculated from the measured rise in borehole water levels and knowledge of the specific yield, throughflow ($2x10^6 m^3$) was derived with the aid of Darcy's Law, evaporation ($78x10^6 m^3$) calculated using the Penman equation and abstractions ($<1x10^6 m^3$) monitored or estimated.

The bulk recharge to the alluvial aquifers was thus $460x10^6 m^3$, or 154mm/y. Infiltration rates for individual wadis, assuming wadi recharge as the major source, ranged from $180,000 m^3/km/y$ to $470,000 m^3/km/y$.

Reference:
HUNTING TECHNICAL SERVICES, 1977. Agricultural development in the Jebbel Marra area: annex 2, volume 1 - hydrogeology

12.7.7 Case 6

This project area is in a particularly arid region of the Middle East, with rainfall ranging from 150mm/y in the upland regions to the west down to <40mm/y in the eastern lowlands. The main aquifer, a 100m thick sandstone sequence, outcrops in the west and any recharge takes place here by infiltration through wadi runoff channels.

The model for calculating recharge allows the first 10mm of daily rainfall to be either evaporated or taken up by depression storage and any subsequent rainfall to become runoff. All the runoff is then assumed to reach a wadi channel from where it either flows out of the area or infiltrates into the wadi alluvium. Surface runoff flowing out of the catchment is assumed to be 4% of the mean rainfall, the remaining runoff then being used to cancel the soil moisture deficit of 200mm in the wadi alluvium before recharge to the underlying aquifer can take place. In an average year the daily rainfall only exceeds 10mm on 0.66 days and results in 4.8mm of runoff. For a catchment of 1800km^2 this amounts to 8.7x10^{6}m^3/y or 10% of catchment rainfall. Of this total, 3.5x10^{6}m^3 flows out of the area and 3.9x10^{6}m^3 cancels the soil moisture deficit, leaving 1.3x10^{6}m^3 to recharge the 90 km^2 bed area of wadi alluvium.

12.7.8 Wadi Jizan, Saudi Arabia

This project area is in the south western corner of Saudi Arabia, and can be considered simply as a 1280km^2 hard rock upland catchment, runoff from which flows along a major wadi channel, crossing a coastal strip of Quaternary alluvial sediments. Some of the wadi flow is diverted to an irrigation project with any remaining surface water eventually flowing to the Red Sea. Average annual rainfall is 480mm for the upland catchment, falling to 150mm on the coastal plain and potential evaporation has been calculated as 3000mm/y using the Penman formula.

The wadi has been gauged and the runoff estimated at 13% of the upper catchment rainfall, giving a mean annual runoff of 80x10^{6}m^3/y. A water balance has been attempted for the plain, using this as the input volume, and possible recharge mechanisms analysed.

The water table, at 6-20m below the surface within the irrigated area, is thought to be too deep to receive recharge from the intermittently flooded surface. For the same reason irrigation canals, which are only used for an average of two days at a time, are not thought to contribute and any percolation losses are eventually evaporated.

Infiltration rates in the wadi bed have been measured at up to 40cm/hr, and as the wadi is subject to longer flow periods it has been taken to be the major recharge source. Wadi recharge has been estimated for the purpose of the water balance at 25x10^{6}m^3/y, which over a distance of 50km results in an infiltration rate of 500,000m^3/km/y.

Reference:
SIR WILLIAM HALCROW & PARTNERS, 1972. Irrigation development in the Wadi Jizan

12.7.9 Wadi Dhamad, Saudi Arabia

The hard rock catchment in this example is adjacent and to the south of that in the last, and the two wadis run along an approximately parallel course to the sea. The upper catchment covers an area of about 1000km^2, for which there is a single raingauge. The mean annual rainfall at this gauge has been estimated at 527mm/y over a period of 14 years, during which time the maximum annual rainfall was 1253mm and the minimum, 209mm. It can be assumed that areal variation in annual rainfall totals within the catchment are similarly extreme, the point being that the mean annual rainfall figure is in all cases only a very approximate guide to the actual catchment rainfall.

The resultant runoff was estimated in the following ways:

(i) using the same runoff coefficient as that for the adjacent project, 0.13, gave a runoff of $69.2 \times 10^6 m^3/y$;

(ii) using data on runoff and catchment area from 9 other catchments to build up a regression equation; this resulted in an estimated average annual wadi flow of $59.7 \times 10^6 m^3$ or 11% of rainfall;

(iii) developing a multiple regression equation, using data from 6 rainfall stations and a gauging station on the adjacent wadi to synthesize the monthly flow records from the gauging station emplaced at the source of this wadi; this gave a mean annual flow of $46 \times 10^6 m^3$ or 9% of rainfall.

Method (iii) is generally considered to be the most rigorous, but ironically the data on which it has been based is from an atypically dry six year period, leading to an underestimate of the mean runoff. This example serves to underline the vulnerability of any estimation technique to the vagaries prevalent in each stage of the recharge process in an arid area.

There has been no direct calculation of recharge from this wadi flow, but the throughput for a 20km wide section of the alluvial aquifer was estimated to be $35 \times 10^6 m^3/y$. This includes the contribution from the adjacent wadi, so assuming a steady state condition, recharge along this wadi is approximately $10 \times 10^6 m^3/y$, or $200{,}000 m^3/km/y$ over 50km.

Reference:
SIR WILLIAM HALCROW & PARTNERS, 1977. Wadi Dhamad irrigation project: volume 2 - main report

12.7.10 United Arab Emirates

This example is from a water resources survey carried out in The Trucial States (now UAE), an arid region of mountains and desert which juts out into the Arabian Gulf from the northern coast of the Arabian peninsula. In common with other Middle Eastern regions the highest rainfall falls on the mountainous inland zone, with wadi outflows being the main source of recharge to the outwash gravels and alluvial plains of the lowlands lying to both the east and west.

The catchment area governing flow to the west has been assessed in some detail; this covers an area of $4{,}000 km^2$ and receives an average annual rainfall of 357mm. Recharge takes place on the gravel plains, cross-cut by wadi systems, which abutt against the mountains' western flanks. To the west lies a sand covered desert plain, and beyond are the coastal sabkhas where discharging groundwater is evaporated. Surface flow from the wadis rarely penetrates any great distance into the desert and as rainfall is low, at a mean of 110mm/y, recharge is thought to be largely confined to the gravels.

A throughflow calculation has been attempted, along a 65km wide front within the desert plain where the transmissivity and groundwater gradient have been assessed from a number of exploratory boreholes. The estimate is then extrapolated across a complete section of the groundwater flow system, resulting in a figure for the total throughflow of $128 \times 10^6 m^3/y$. Abstractions up-gradient of this section account for a further $70 \times 10^6 m^3/y$, so assuming the system is in equilibrium and wadi flow from the upper catchment is the only recharge source, 14% of the upper catchment rainfall can be said to recharge the groundwater system.

Reference:
SIR WILLIAM HALCROW & PARTNERS, 1967. Water resources survey of The Trucial States: volume 2 - draught report

13 INTERAQUIFER FLOWS

Flows to and from other aquifers can be an important part of the water balance of an aquifer, and must be taken into account in modelling. Usually seepage is from an underlying aquifer with low flows and reasonably steady conditions. Although seepage may be small when expressed per unit area, large areas are often involved.

Using Darcy's Law can give good estimates and is an approach easily linked to modelling. It is usually possible to measure hydraulic heads in both aquifers; the major uncertainty is knowing the hydraulic conductivity of the intervening material. Laboratory measured values are rarely related to regionally useful field values. A common approach in modelling is to set up a leaky boundary, and to include the conductivity as one of the parameters to be calibrated.

Water balances of both the donating and receiving aquifers should always be used to set limits on the likely interaquifer flow. More accurate results will be obtained when the interaquifer flow is a large proportion of the water balance. The direction of flow can be determined by head measurements, hydrochemical or isotope data, or from preliminary numerical modelling of the system.

Tracer techniques, particularly isotopes, are useful in detecting flow between aquifers. The main condition is, of course, that the isotopic composition of each aquifer's water is significantly different, say by several parts per mill in the case of oxygen-18. However, tracer methods rarely give firm estimates of the amount of inter-aquifer flow, because of the unknown mixing of the two waters in any sample. There may be a role for transit time calculations in steady state (or small) groundwater systems. More information is given in Chapter 21 of the Guidebook on Nuclear Techniques (IAEA, 1983).

14 RECHARGE FROM IRRIGATION
(with contributions by Robert Gray)

14.1 Introduction

14.1.1 Importance of irrigation recharge

Irrigation schemes are frequently a major source of recharge to aquifers, whether they use groundwater or surface water. For example, Lerner et al. (1982) estimated that the alluvial fan aquifer under Lima, Peru received 20% of its recharge from irrigation losses. There have been many studies of groundwater in the Indus Basin where groundwater levels rose at least 20m in the 30 years after irrigation began (Greenman et al., 1967).

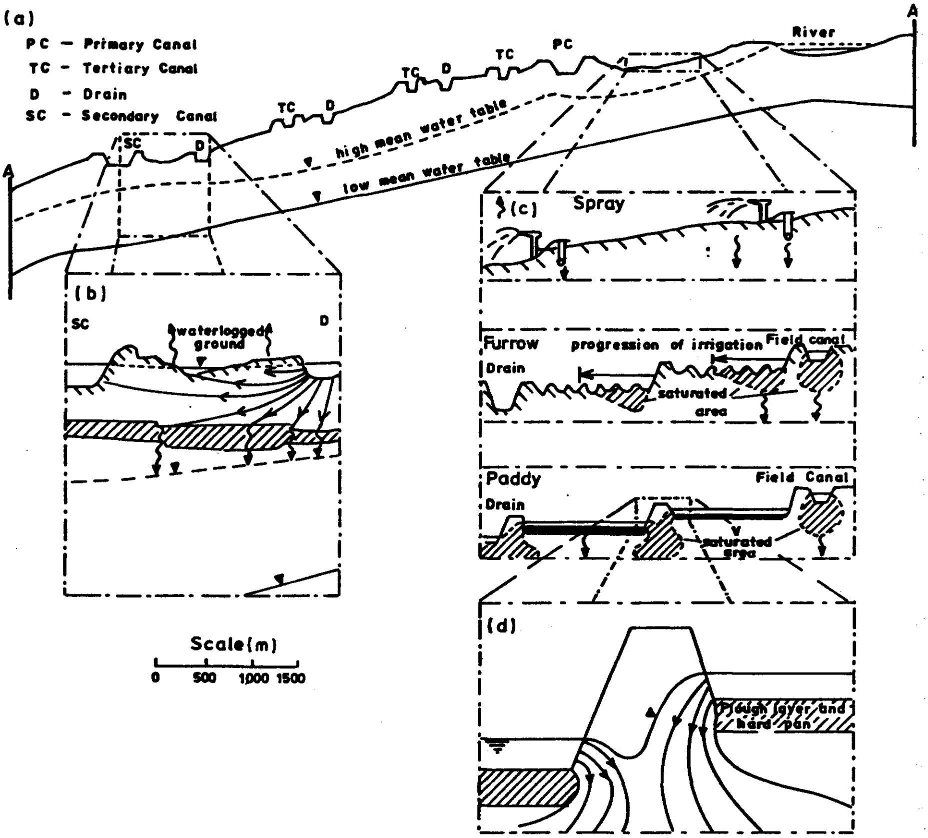


Fig. 14.1 Cross-section of a hypothetical irrigation scheme showing potential recharge sources (a) cross-section with arrangement of river, canals and drains and showing two areas enlarged below (b) waterlogging caused by low permeability layer (c) spray, furrow and paddy irrigation (d) enlargement of paddy bund showing high seepage losses

Fig. 14.1 shows a schematic section through a hypothetical irrigation scheme. Losses from irrigation, which are potential recharge as defined in Section 11.1.2, occur in two overlapping areas (a) canals and (b) fields. Canal losses are probably easier to estimate, and have received more attention. The bulk of this chapter (Sections 14.2-6) reviews these methods.

Field losses will vary with the type of irrigation: flooded paddy, intermittently flooded field, furrow or spray irrigation. Drip irrigation is not considered here; if there are any losses they will probably be measured by the operators. Field losses are considered in Section 14.7.

Recharge from irrigation schemes is sometimes underestimated because it is neglected. On the other hand, once irrigation has been recognised as a recharge source, the quantities are often overestimated because all losses are assumed to become recharge. The difference between potential and actual recharge is discussed in Section 11.1.2. Rushton (1986, Chapter 23) and WRRC (1980) consider the issue in more detail for irrigation, and Rushton (Chapter 23) gives a good example of the difference in the Mehsana aquifer in Gujarat, India, where the water table is 20m below ground. Substantial losses of water were observed from the canal irrigation scheme which has recently spread over the aquifer, and groundwater levels were expected to rise in response. This has not occurred. Instead waterlogging is now seen along many canals; the losses have formed a perched water table on top of low conductivity layers in the aquifer, and are evaporating or being taken away by drains.

A general procedure for estimating irrigation recharge is discussed in Chapter 10, and is particularly important for irrigation because the high potential rates often exceed the ability of the aquifer to accept, or of the unsaturated zone to transit, the water.

If actual recharge is substantially less than potential, then care is needed in building a regional groundwater model. Actual recharge may well depend on groundwater levels, and these may alter significantly in the future. The modelling aspects of this problem are discussed in Chapter 10.

14.1.2 Water losses from irrigation canals

The loss of water from canals can amount to a significant percentage of the total irrigation supply. This often results in the waterlogging and consequent abandonment of surrounding arable land, as well as representing a substantial depletion of the water resource. Case studies (Ahmad 1974; Worstell, 1976; Pontin et al., 1979) reveal canal losses which, when expressed as a loss over the wetted area of the bed, vary from 0.05 m/d to 1.5 m/d. Kraatz (1977) lists the estimated water losses from 15 canal systems around the world; seepage losses ranged from 3-50% of the water carried.

Total water loss from a canal is the sum of evaporative and seepage losses. Losses due to evaporation can be calculated by the Penman equation and will probably be in the range of 1-10 mm/d. Seepage losses may either flow to shallow water tables and thence to drains and evaporation, or recharge deep groundwater systems.

Canals are often lined in an attempt to cut down on losses. Rushton (1986) argues that even small imperfections in a lining allow large losses. He bases this on a numerical model study in which 0.4% of the lining was assumed to be missing; water losses were still 67% of those from the unlined model.

Figures taken from the Periyar Vaigi Project in South India (Rushton, 1986) and reproduced in Table 14.1 illustrate the relative ineffectiveness of canal lining in that region, and Table 14.2 gives an example of the reduction in effectiveness due to a deterioration in the condition of the lining.

Table 14.1 An example of the effect of canal lining on seepage losses, Periyar Vaigi Project, South India (Rushton, 1986).

Type of canal	**Seepage losses (m/d)**	
	Unlined	**Lined**
Main canal	0.37	0.11
Large distributary	0.18	0.08
Medium distributary	0.09	0.06
Small distributary	0.06	0.045

Table 14.2 The effect of deterioration in lining condition on seepage losses, Salt River Project, Arizona, USA (WRRC, 1980).

Condition of lining	**Seepage losses (m/d)**
Unlined	0.25
Concrete lined, fair condition	0.07
Concrete lined, good condition	0.003

There are many approaches to the calculation of canal losses, as detailed below, but most of these contain assumptions that may not be valid for the canal under study. Two of the main assumptions are:

(i) The canal and groundwater are in a steady state condition. This is rarely the case as agricultural needs and water supplies dictate that water levels are constantly changing, disrupting any steady state interaction with the underlying strata.

(ii) The strata beneath the canal can be simplified into a one or two layer model. Variations in permeability with depth are shown by Rushton (1986) and Wachyan & Rushton (1987) to have a profound effect on the eventual recharge. This complexity is however beyond the scope of most of the theoretical models presented here.

14.1.3 Methods of estimating recharge from irrigation canals

As with most recharge calculations, there is not a perfect method for all, or even individual cases. Direct measurement has been used for seepage rates through canal beds (Section 14.2). Empirical formulae are perhaps more widely used for this type of recharge than any other, but most formulae have been developed to estimate canal losses for design and operation purposes (Section 14.3). Water balances are commonly and successfully used to estimate canal losses (Section 14.4), and there are a number of tracer techniques but these have not been widely used (Section 14.5). Darcian approaches include flow nets, analytical solutions to the flow equations, and numerical models (Section 14.6). Table 14.3 briefly compares these groups of methods.

14.2 Direct measurement of canal losses

Seepage meters have been developed to measure canal losses. They involve a seepage bell or cylinder which is pushed into the sediment at the base of the canal and the infiltration rate measured by constant or falling head techniques. Some of the advantages to their use are as follows:

- lightweight and easily transported,
- relatively cheap,
- simple to operate,
- rapid measurement,
- straightforward conversion to a seepage value.

Of course the base of the canal must be sufficiently soft to insert the cylinder. This rules out its use on concrete or rock lined canals. Difficulties may also be encountered in gravelly or stony soils, and sandy soils may be washed from around the seepage cylinder by eddy currents. The soil can be disturbed by insertion, giving higher infiltration rates. Comparison of results from seepage metering and the ponding method (Section 14.4.2) has shown 30% higher seepage losses recorded by the seepage meter (Brockway & Worstell, 1968). Warnick (US Agric. Research Service, 1963) also compared two methods: pushing the seepage bell into the sediment gave 23% higher rates, and hammering the bell into position resulted in 52% higher seepage rates than those recorded by the ponding technique.

The number of measurements taken per unit area to arrive at a reasonable average is dependent on the degree of heterogeneity in the seepage loss at the specific site. Brockway & Worstell (1968) working in Arizona, made one seepage measurement for every 100m^2 of canal base. Two experienced men were able to perform about 40 tests per day. The cost, at $180 per km on

Table 14.3 Comparison of methods of estimating recharge from irrigation canals

This table is not a substitute for the longer discussion of the methods given in the main text of this chapter.

Method	Direct measurement	Empirical formulae	Water balance	Borehole tracer methods	Analytical Darcian methods
Section	14.2	14.3	14.4	14.5.2-14.5.3	14.6.1-14.6.2
Applicability	Soft bedded canals of low to moderate velocity	Not recommended except as first estimate of losses in similar terrain as that used to derive formula	Seepage losses from most canals	Recharge from canals in continuity with groundwater	Theoretical and parametric studies. Recharge from canals in homogeneous ground under steady state conditions
Accuracy	Reasonable	Poor, especially for recharge	Moderate to good	Poor	Variable, usually poor
Data requirements	All collected in the field	Varies with formulae used, usually small	Either in and out flows or water levels and dimensions of canal	Collected on site	Hydrogeological conceptual model, permeabilities
Ease of use	Straightforward	Straightforward	Straightforward but requires field work with attendant logistical problems	Moderate, once boreholes drilled and procedure established	Simple if published result used, difficult if new solution developed
Type of estimate	Instantaneous point estimate of seepage rate	Usually average loss per kilometre	Average seepage loss over period of measurement	Average (unknown time period) recharge in a cross-section	Steady state canal losses from cross-section
Costs	Low	Low	Moderate to high for fieldwork	High if boreholes to be drilled	Low
Time required	Short	Short	Moderate: one time survey is quick but not very accurate	Short once boreholes established	Short to moderate

an 8m wide canal, compared favourably with the ponding method at $900 per km.

In conclusion, the seepage meter gives a rapid and direct measurement at low cost, but the figures obtained are only point measurements and are therefore not best suited to estimating seepage losses for the whole canal. More appropriate uses for the meter are, for example, evaluating the effectiveness of a penetrable lining material prior to its general use, or the location of areas of heavy seepage along a canal length.

Constant head seepage meter. This is the simpler of the two common types of meter and is illustrated in Fig. 14.2. The cylinder is inserted while the hose is left open to allow air and excess water to escape. The plastic bag is attached to the hose, and the bag allowed to float just below the surface. This should result in an equal pressure head to drive the seepage both inside and outside the cylinder. The bag's volume is measured again after a time has elapsed and the seepage rate is calculated from

$$q = (V_1 - V_2)/(\pi\ r^2\ t) \qquad 14.1$$

where q : seepage rate (L/T)
V_1 : initial volume (L^3)
V_2 : volume after time t (L^3)
r : internal radius of the seepage cylinder (L)

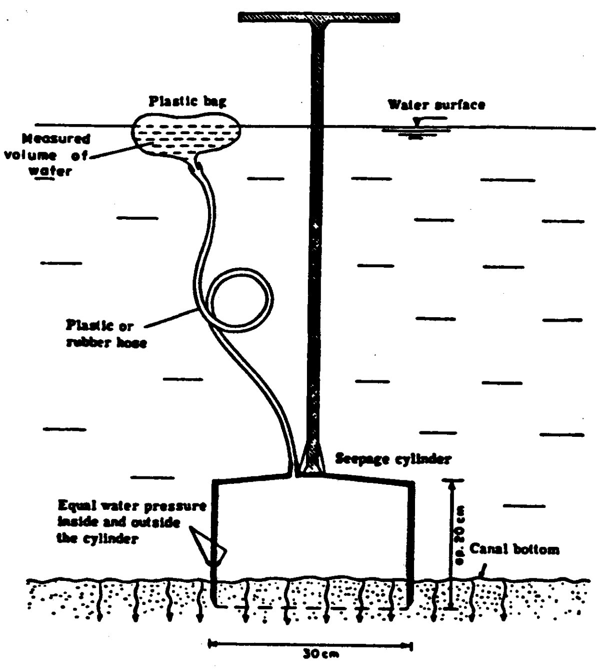


Fig. 14.2 Constant head seepage meter for canal bed (Kraatz, 1977, Fig. 14)

The major difficulty with this method is the preservation of an equal internal and external head. The effect of the variation in head increases for decreasing seepage rates; a head difference of 1 cm can cause measurement errors of the same order as the seepage for low seepage rates (US Agricultural Research Service, 1963).

Falling head seepage meter. Bouwer and Rice (1963) describe the design and use of a falling head seepage meter.

14.3 Empirical methods for canals

There are many empirical formulae used to estimate canal loss - that is potential recharge. Few of the formulae are designed to estimate actual recharge. The formulae are usually based on prolonged observation of canals within a region, and as such will not be valid for canals in other regions with different field conditions (Krishnamurthy & Rao, 1969).

In general, empirical formulae relate seepage losses to one or more of the following:

(i) the soil type and condition,

(ii) the depth or flow of water in the canal,

(iii) the wetted perimeter.

This last is seen in many empirical formulae, but Rushton (1986) argues that it is not a controlling factor in most cases. It is only appropriate when a canal is high above the water table so that recharge travels vertically downwards. This may occur when a canal starts to carry water at the beginning of the irrigation season and the soils below are dry.

The following features can also have a profound effect on the seepage loss but are rarely considered in empirical formulae:

(iv) the depth to the regional water table,

(v) layering within the soil,

(vi) clogging of the canal base with fine sediment,

(vii) irregular use of the canal, and the consequent variation between saturated and unsaturated conditions.

Other authors have cited the following parameters as being major factors:

(viii) cross-sectional geometry,

(ix) canal discharge velocity.

Kraatz (1977) lists six formulae used in various parts of the world. Three examples are given in the following paragraphs, with some corrections and changes to units.

The International Commission on Irrigation and Drainage (1968) quote the following equation, derived from observation of several canals in the Punjab, India:

$$S = c\ a\ d \tag{14.2}$$

where S : total seepage loss (m^3/s)
c : a constant found to vary from 1.1 to 1.8 for those canals observed
a : area of wetted perimeter (km^2)
d : canal water depth (m)

Offengenden (FAO/Unesco, 1967) proposes the following equation for estimating water loss from earth canals or ditches:

$$S = A.L/(100.Q^{m-1}) \tag{14.3}$$

where S : total seepage loss (m^3/s)
Q : discharge (m^3/s)
L : canal length (km)
A,m : empirical constants dependent on soil permeability as follows:

	Permeability		
	Low	Medium	High
A	0.7	1.9	3.4
m	0.3	0.4	0.5

Egypt's Irrigation Department use the formula devised by Molesworth and Yennidumia (Doorenbos, 1963):

$$S = b\ P\ L\ R^{1/2} \tag{14.4}$$

where S : total seepage loss (m^3/s)
L : length of canal (km)
P : wetted perimeter (m)
R : hydraulic mean depth (m)
b : soil coefficient, varying from 0.0015 for clay to 0.003 for sand

An empirical formula developed for a particular region and used in conjunction with field data for that region, may well constitute a valid corroboration of the field evidence. The formula should still however be regarded as a simplification of the seepage process, and formulae from one region are therefore likely to be totally inappropriate elsewhere.

14.4 Water balances on canals

14.4.1 Inflow-outflow measurement

The canal is gauged at two or more points along its length, with the distance between gauged points varying from a few hundred metres for small flows up to several kilometres for larger flows. Once any intermediate inflows and outflows have been accounted for, the shortfall in the output is attributed to seepage loss.

This method is well suited to small canals of simple design and constant flow, particularly where seepage losses are thought to be large and show considerable spatial variation. Where discharge varies, intakes or offtakes are numerous, or where the flows are very large in relation to the seepage loss, the measurement error may well be of the same order as the seepage, making analysis difficult. The US Bureau of Reclamation (1967), Starosolszky (1959) and Scott & Houston (1959) contain more detailed information on the use of this method.

Flows can be measured by current metering or other velocity methods which use floats or tracers, see eg WMO (1980) or Herschy (1978). These methods are cheap and require no structure, but only provide instantaneous values. Purpose built structures equipped with automatic recorders, provide the most accurate measurements. These are rarely available on canals, but canal control structures can be calibrated as flow measurement devices, see eg Bos (1978) or Tilrem (1986).

The advantages of the inflow-outflow method are as follows:

- both the practical and theoretical considerations are straightforward and rely on the minimum of assumptions.

- The calculated seepage loss is an average figure over the area of the test.

And the disadvantages:

- all intermediate inflows and outflows have to be monitored.

- Where the flow varies with time (a) a lag time has to be calculated between the measurement of the upstream and downstream flow and (b) errors are introduced due to the effects of bank storage.

14.4.2 Ponding loss measurement

A section of the canal is isolated and the rate of fall in water level is measured. When evaporation losses and any rainfall input have been accounted for the net fall in water level is attributed to seepage. The rate of seepage loss per unit length is then calculated:

$$q = (d_1 - d_2)\ (W_1 + W_2)/(2\ t) \qquad 14.5$$

where q : seepage rate per unit length ($L^3/T/L$)
d_1 : initial mean depth (L)
d_2 : mean depth after time t (L)
W_1 : initial mean width (L)
W_2 : mean width after time t (L)

The construction of dams to isolate a canal section, and their subsequent dismantling, may prove to be relatively costly. Kraatz (1977) reports the use of the ponding method in western Macedonia, Greece. Jute sacks filled with loam soil were used to construct the dams, and watertightness was ensured by packing layers of the soil between each sack layer. A stake driven into the canal base served to measure the fall in water levels.

The damming of a canal section may reduce seepage losses because fine sediment settles and algae may start to flourish after several days. This effect is most apparent on those canals that are normally rapid flowing. The reduction in seepage loss can be accounted for by plotting the fall in canal level at regular time intervals, as shown in Fig. 14.3. The tangent to the early time data represents the corrected depth-time relation, and d_{2c} the corrected depth.

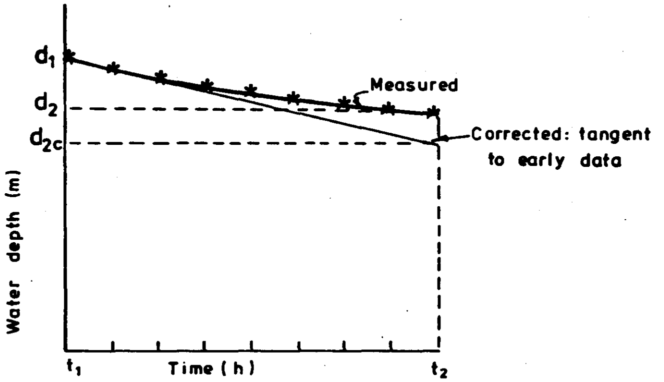


Fig. 14.3 Depth correction for ponding loss method

As the canal level falls, (a) water is released from bank storage, and (b) the head difference between the water in the canal and the regional water table also falls. Both these factors reduce losses relative to those when the canal is flowing. Constantly topping up the canal to its initial height with measured volumes of water overcomes this potential error.

The method can be recommended as an accurate field technique and compares well with inflow-outflow, particularly when seepage losses are small. It is however less convenient as the canal is out of service for several days, and more expensive than the use of seepage meters. Further information can be obtained from Brockway & Worstell (1968) and Correia (1963).

14.5 Tracer techniques for canal seepage

Tracer techniques have some application to the analysis of seepage losses. The following methods are mentioned in the literature:

- bed penetration,
- point dilution,
- two well technique.

The bed penetration technique was developed by Bouwer & Rice (1968) specifically for the measurement of seepage loss. It allows a number of point measurements to be made over a limited area. In contrast, the point dilution and two well techniques are better known as methods of assessing the filtration velocity, as part of a regional hydrogeological investigation. In this adaptation the filtration velocity is calculated along a plane normal to the canal axis and the seepage outflow estimated with the help of Darcy's Law.

14.5.1 Bed penetration

This technique was originally developed to use salt as a tracer. It was spread on the canal bed as crystals and left. An electrical conductivity probe was then inserted into the canal sediment and the depth of the peak conductivity reading denoted the distance travelled by the seepage water during the intervening period, from which the seepage rate can be calculated.

The salt application required to achieve a pronounced conductivity peak is in the range of 1-5kg/m^2. Application from a boat, which will not cause a major disturbance to the canal sediments, is preferable to wading. Where the canal has a clay base, the sodium ion in sodium chloride will react with the clay, causing deflocculation and a consequent reduction in hydraulic conductivity. A salt which will not react, such as calcium chloride or aluminium sulphate, should be used.

It may be possible to use other applied tracers, such as radioisotopes, which can be detected in situ at low concentrations.

The conductivity/depth plots should be of the same form as one of the curves shown on Fig. 14.4. Curve A represents the situation where there has been rapid dissolution of the salt, after which low salinity canal waters re-enter the sediment. If salt crystals are present on the canal bottom for some extended period of time curve B results, whereas curve C shows the readings to be expected if the salt crystals persist on the surface until the electrical conductivity probe is inserted. The points P_A and $P_{B,C}$, the characteristic points, represent the position of the saline seepage front for use with the corresponding curves and the depth to be used for seepage rate calculations. However, if the curve shows the highest conductivity at the surface, falling off rapidly with depth (curve D), then no seepage has taken place.

The seepage rate is obtained from the following

$$q = (d\ p)/t \qquad 14.6$$

where q : seepage rate (L/T)
p : porosity
d : depth of point P at time t (L)

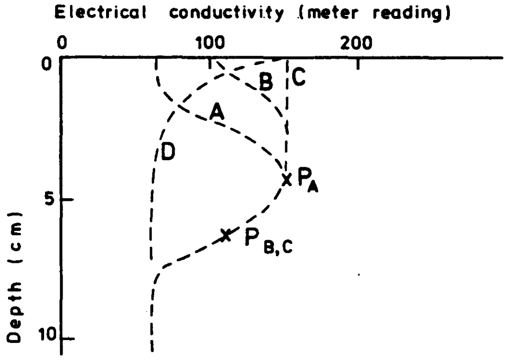


Fig. 14.4 Possible conductivity profiles using the salt penetration technique

Bouwer & Rice (1968) present a detailed appraisal of the technique. They found that for an accurate estimate of seepage rate, the point P should be allowed to advance from 5 to 15cm into the underlying sediment, beyond which depth adsorption and diffusion effects make the peak more difficult to detect. For a seepage rate of 2m/d this position will have been reached after 30 minutes, whereas at rates of 5 to 10cm/d sixteen hours would have to elapse. It is therefore recommended to take probe readings at 30 minute intervals until the point P is at a suitable depth, at which time a more detailed analysis can be made.

14.5.2 Point dilution: single well method

This method aims to calculate the filtration (pore water) velocity away from the canal, by adding a known concentration of tracer to a borehole positioned within the adjacent strata and then monitoring the rate of dilution of this tracer (Halevy et al., 1967). Tracers which can be measured in situ are usually prefered. Salts, such as sodium or calcium chloride, can be used when the salinity of the water is low. The concentration can be estimated by measuring the water conductivity. Radioactive tracers are very suitable because they can easily be measured at very low concentrations, avoiding density induced flows.

Krishnamurthy & Rao (1969) used Bromium-82, in the form of potassium bromide, and Iodine-131, as potassium iodide, as tracers to estimate seepage from the Ganga canal. These two are the most commonly used radioactive tracers for this purpose.

A tracer is placed in a borehole and its concentration measured over time. An initially well mixed tracer can be created by forcing air from a pipe lowered to the bottom of the borehole, or moving a wire spiral slowly up and down within the measuring volume. The filtration velocity can be calculated from:

$$v_f = [\pi\ r_1\ \ln(c_0/c_t)]/(2\ a\ t) \qquad 14.7$$

where v_f : filtration velocity (L/T)
r_1 : internal radius of the well screen (L)
c_0 : original tracer concentration
c_t : tracer concentration after time t
a : correction factor to allow for hydrodynamic distortion

Kol (quoted by Halevy et al., 1967) established theoretically that a is approximately 2 for a simple uncased borehole. Where the borehole is enclosed by a gravel pack (Fig. 14.5a, the relation between "a" and the pack is given by (Drost, 1981):

$$a = 8\ K_2/(A\ K_2\ +\ B\ K_1) \qquad 14.8a$$

where A and B have approximately the same numerical value and

$$A = (1+K_2/K_1)[1+(r_1/r_3)^2] + (1-K_2/K_1)[(r_1/r_2)^2 + (r_2/r_3)^2] \qquad 14.8b$$

$$B = (1+K_2/K_1)[1-(r_1/r_3)^2] + (1-K_2/K_1)[(r_1/r_2)^2 - (r_2/r_3)^2] \qquad 14.8c$$

where r_1 : internal radius of the well screen (L)
r_2 : outer radius of the well screen (L)
r_3 : radius of drilling (L)
K_1 : hydraulic conductivity of the well screen (L/T)
K_2 : hydraulic conductivity of the gravel pack (L/T)

The validity of eqns 14.7 and 14.8 is based on the assumptions that the flow is steady state, the tracer is perfectly mixed within the borehole, there are no vertical flow components within the borehole, density currents are not present, and diffusion effects can be ignored.

For the purposes of this experiment, flow can be assumed to be steady-state if the canal is in constant use and there is no great variation in discharge. The velocity at which turbulent flow invalidates the method varies with well construction in the range 20-300 m/d. Diffusion becomes important at low velocities and should be taken into account for v_f < 0.005 m/d. To avoid density currents the concentration should be kept below 4.10^{-4} mol/l and injection carried out after it has reached borehole temperature.

In a single aquifer, the labelling of the total water column, and recording of later concentration profiles will give a vertical profile of the filtration velocity (Plata-Bedmar, 1983).

In a heterogeneous section pressure gradients often build up between the different permeability strata. The open path produced by a borehole provides a short circuit between these layers, resulting in vertical flows within the borehole which further dilutes the tracer. Measuring these vertical, usually downward, velocities may provide useful qualitative information on hydrogeology, but the use of packers to isolate sections of similar permeability is necessary to quantify horizontal velocities.

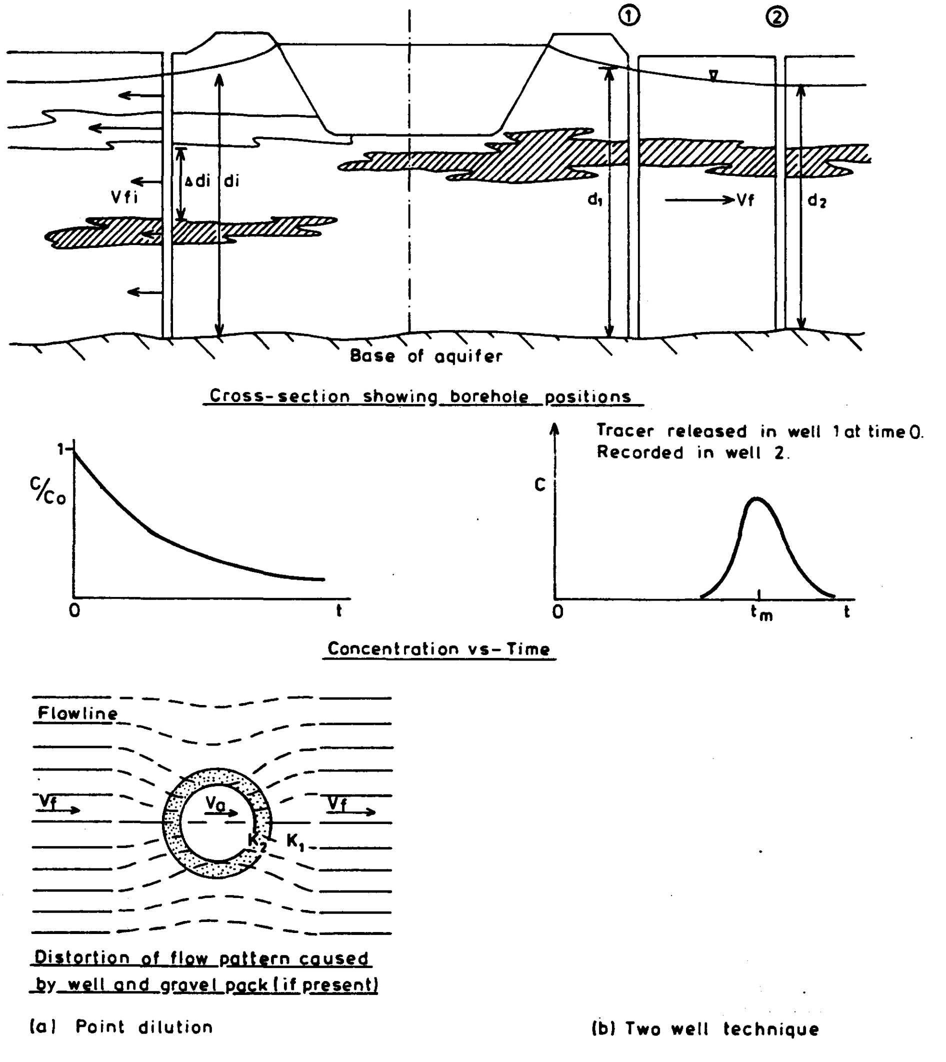


Fig. 14.5 Borehole tracer techniques for filtration velocity and canal seepage (a) point dilution with single borehole (b) two hole technique

The inaccuracies inherent in this method of calculating filtration velocity are such that, after all these corrections have been made to the field results they still cannot be expected to give better than ± 50 % accuracy for absolute values of v_f (Halevy et al., 1967).

Fig. 14.5 shows a schematic cross-section of a canal and the variables involved in tracer methods. The filtration velocity is measured at depth intervals over the full depth of canal seepage influence, resulting in a velocity profile down the borehole. The rate of seepage loss per unit length of the canal is given by

$$q = 2 \Sigma v_{fi} \ d_i \qquad 14.9$$

where q : seepage rate per unit length of canal (L^2/T)
v_{fi} : filtration velocity at measurement point i (L/T)
d_i : length of borehole over which v_{fi} applies (L)

The multiplier of 2 is needed if the experiment is conducted on one side of the canal. However, it is advisable to conduct this test on both sides of the canal to allow for the effects of asymmetry in the seepage losses and regional groundwater gradients. Krishnamurthy & Rao (1969)in their study of seepage losses from the Ganga Canal measured seepage losses of 11mm/d and 19mm/d for the right and left hand banks.

Other authors have used more complex methods to estimate the seepage losses, (eg Krishnamurthy & Rao, 1969), but although these methods appear more mathematically precise, they tend to introduce further assumptions and result in no more accurate a solution than the above.

14.5.3 Two well method

In this method a tracer is introduced into the injection borehole and its subsequent arrival is monitored at another borehole down gradient and close enough to give a reasonable transit time. The breakthrough curve (plot of tracer concentration against time) is plotted, and the mean transit time is used to calculate the filtration velocity

$$v_f = L/t_m \ p \qquad 14.10$$

where v_f : filtration velocity (L/T)
L : borehole separation (L)
t_m : mean transit time (T)
p : porosity

The filtration velocity can then be used to calculate the seepage flow, from eqn 14.9.

This method is limited to regions with a relatively high filtration velocity, and its accuracy is dependent on the accuracy of the transit time calculation. This, in turn, is reliant on a simple breakthrough curve. Field tests carried out within a heterogeneous aquifer may result in a breakthrough curve consisting of more than one peak, making analysis more difficult.

14.6 Darcian approaches for canal seepage

14.6.1 Introduction

Canals have many similarities with rivers, and so the general points of the discussion of Darcian approaches for rivers (Section 12.4) apply to canals, but with differences of emphasis and detail. Four groups of Darcian approach have been used for canals:

(i) infiltration equations and flow nets,

(ii) gathering field data on aquifer properties and piezometric heads,

(iii) analytical solutions to the equation of flow, usually in homogeneous and isotropic aquifers,

(iv) analogue or numerical models.

Most of the Darcian techniques employ one of a limited number of conceptual models. The first four models were concisely presented by Bouwer (1965), the others by Wachyan & Rushton (1987) as being more realistic of field situations. They are shown in Fig. 14.6 and are:

(a) The soil in which the canal is embedded is uniform and underlain by more permeable (considered infinitely permeable) material. The regional water table is above the top of the permeable layer. (Bouwer's condition A).

(b) The soil in which the canal is embedded is uniform and underlain by more permeable material (as above), but the water table is at or below the top of the permeable layer (Bouwer's condition A').

(c) The soil in which the canal is embedded is uniform and underlain by impermeable material. Canals alternate with drains, giving a fixed head drainage boundary on both sides (Bouwer's condition B).

(d) The perimeter of the canal is of much lower conductivity than the rest of the aquifer, for example it is lined. This can be associated with any of the above conditions (Bouwer's condition C).

(e) The aquifer is assumed to be underlain by a thin layer of lower hydraulic conductivity, which in turn is underlain by more permeable aquifer material in which the main water table is located. The canal is one of an equally spaced series.

(f) As case (e), but with water withdrawn from the upper aquifer across the water table. This represents

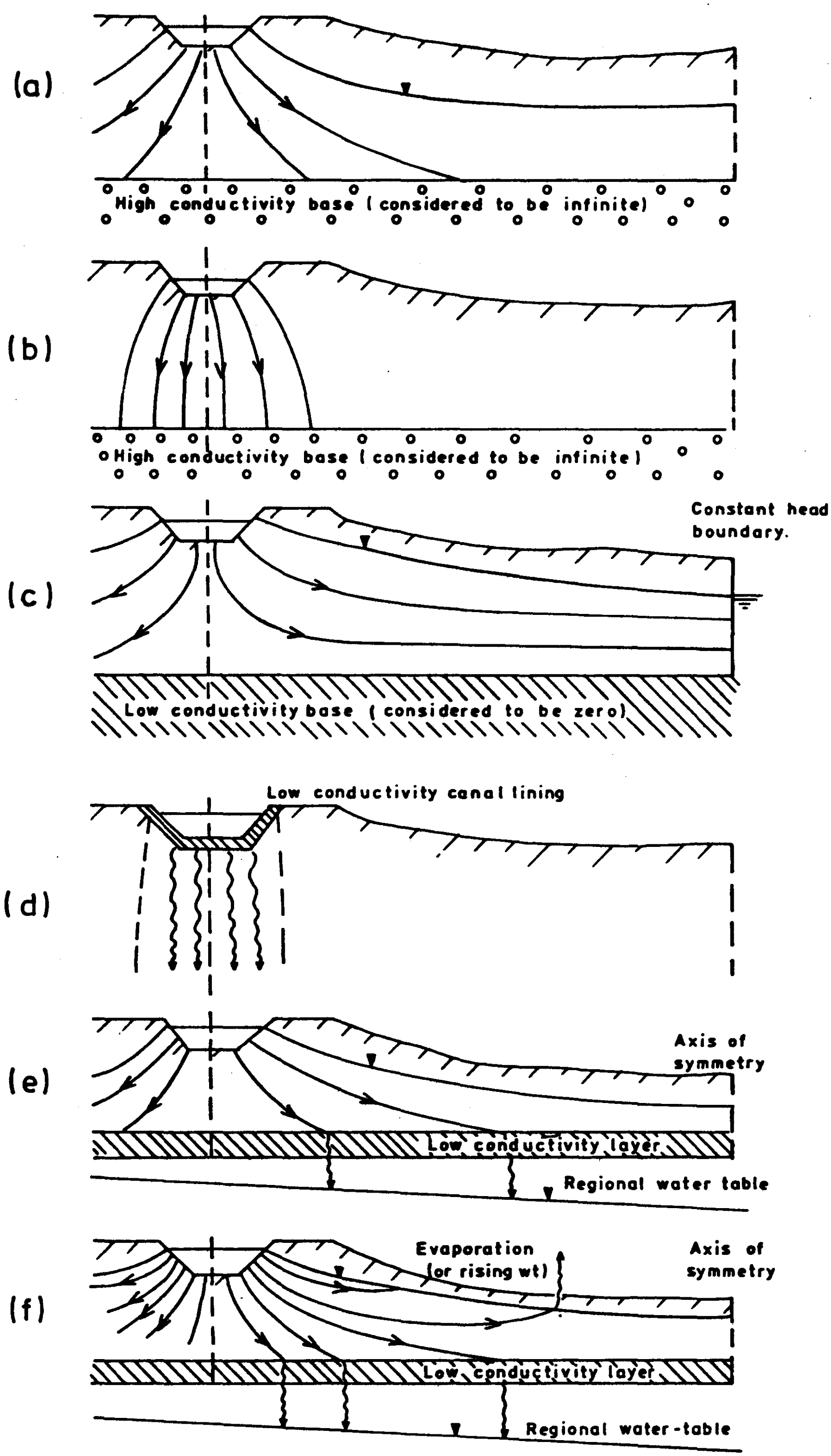


Fig. 14.6 Conceptual models for canal seepage used for many modelling studies

either evapotranspiration, a rising water table, or waterlogging and drainage.

Wachyan and Rushton (1987) show that the apparently small changes between cases (c) and (e) doubles the seepage loss. The difference between cases (e) and (f), ie allowing evapotranspiration can almost double the loss again.

Infiltration equations and flow nets are discussed in Section 12.4.1. The use of field data is briefly described in Section 12.4.2; it is often more applicable to canals than rivers because of the more constant conditions. Analytical solutions are described in Section 14.6.2 below, and a brief review of numerical models is given in Section 14.6.3.

14.6.2 Analytical methods

These methods find an analytical solution to heads and flows. Results are often presented as formulae or graphical relationships. Each analytical method has a restricted range of operation; many have further simplifications to the models shown in Fig. 14.6. The variety and complexity of many of these methods precludes their detailed analysis in this manual, and we only describe the range of situations for which they are viable.

The Dupuit-Forchheimer equation (De Wiest, 1965) is based on the assumption of horizontal flow, but can be used to estimate losses from a canal to a shallow water table in case (c). Ideally, the canal should also be rectangular and fully penetrate the permeable upper layer, in which case the seepage loss per unit length can be approximated as follows:

$$q = [2\ k\ D_w\ (h - D_w/2)]/(L - d/2) \qquad 14.11$$

(Notation with equation 14.12.)

This equation can be adapted for use on a trapezoidal canal with the impermeable layer at some depth below the canal base:

$$q = [2\ k\ D_w\ (h + D_i - D_w/2)]/[L - (d + B)/4] \qquad 14.12$$

where D_w : vertical distance between water level in the canal and the phreatic surface, at a horizontal distance L from the centre line of the canal (L)
- B : width of the canal at the water surface (L)
- h : depth of water in the canal (L)
- L : see D_w (L)
- d : width of the canal at its base (L)
- k : hydraulic conductivity of the layer (L/T)
- D_i : vertical distance between canal bottom and impermeable layer (L)

The error in q increases with increasing D_i but reasonably accurate results can be obtained for values of D_i less than 3 times the width (Mahajan, 1985).

Many authors have tackled the problem of seepage from a canal in a homogenous and isotropic medium in case (b) above. This is the easiest case because the regional water table does not influence the seepage losses; unfortunately it is rarely found in practice. Examples include:

(i) Vedernikov (1936) gave solutions for both triangular and trapezoidal cross-sections (reproduced in Muskat, 1946 and Harr, 1962),

(ii) El Nimr (1963) produced further solutions for a triangular cross-section, subsequently improved by Bruch & Street (1967),

(iii) Morel-Seytoux (1964) gave a solution for a rectangular channel section,

(iv) Kozeny (1933) analysed losses from a curvilinear example.

The following analytical methods all include the influence of a regional water table:

(v) Garg & Chawla (1970) obtained a solution for a trapezoidal canal in a homogenous and isotropic medium with a water table at finite depth and flow lines to both horizontal boundaries, and vertical drainage. This is intermediate between cases (c) and (e).

(vi) Dachler (1933) presented a semi-empirical method of estimating seepage for a similar section but with an impermeable layer at depth, case (c). Sharma & Chawla (1970) developed an analytical solution to the same problem using conformable mapping techniques.

(vii) Hammad (1960) treated the case where the layer at depth is more permeable than the surface layer, case (b).

(viii) Bouwer (1965) developed an analytical solution for a canal where the hydraulic conductivity of a thin layer around the perimeter of the canal is much lower than the conductivity of the original soil, case (d).

14.6.3 Modelling techniques

The situations shown in Fig. 14.6 can be solved by numerical modelling techniques, as can more complex and field based sets of boundary conditions. They have of course many similarities with river models, which are discussed in Section 12.4.3. Specific modelling studies of canals include Bouwer (1964) and Wachyan and Rushton (1987). Detailed descriptions of methods are beyond the scope of this manual, but can be found in a number of standard texts including Rushton and Redshaw (1979), Wang and Anderson (1982) and Huyakorn and Pinder (1983). Such modelling studies can be very useful to understand canal-groundwater interactions; if sufficient data are available, they can help to quantify recharge.

14.7 Recharge from irrigated fields

Recharge by deep percolation from irrigated fields has many similarities with recharge by precipitation, and the same methods can often be used to estimate it. It is an areal process, water is applied at intervals, some may run off, the remainder becomes soil moisture where evapotranspiration forces are applied before the excess becomes recharge. However there are differences between irrigation and recharge which may be important in individual cases:

(i) field canals may be present, carrying water at all times;

(ii) in flooded (paddy) irrigation, substantial depths of water are held on the surface leading to saturated conditions in at least the upper part of the soil profile;

(iii) soils are commonly puddled in flooded irrigation to reduce infiltration rates, however the bunds between fields may be a highly permeable pathway for deep percolation (Walker & Rushton, 1984).

Provided account is taken of such special circumstances, irrigated field recharge can be estimated by the same methods as used for precipitation recharge which are outlined in Chapter 11.

Direct measurement. Lysimeters, described in Section 11.2 can be used in principle. There are obvious difficulties in cultivating the surface to be representative of the surrounding fields which may make their use impractical.

Water balance methods. Soil moisture budgetting methods are discussed in Section 11.4. Although they cannot always be applied for precipitation alone in arid and semi-arid climates, they are more likely to be valid when there is a regular supply of irrigation water.

Inflow-outflow balances of irrigated areas have sometimes proved successful in estimating net recharge. These will take account of canal inflows, precipitation and other water sources, while outflows to be measured include drainage and crop water use; the last is the most difficult.

Darcian approaches. The Darcian approaches outlined in Section 11.5 (numerical modelling, field data) are equally applicable to irrigated fields, and may give more accurate results because of the more regular conditions.

Tracers. Tracer methods are commonly used for estimating recharge from irrigated fields. Methods are described in

Section 11.6. Signature methods using applied tracers are the most common.

14.8 Examples of project studies of irrigation recharge

14.8.1 Introduction

It is often necessary, as a result of the large seepage losses from canal fed irrigation systems, to design and build a drainage network to cope with the waterlogging and consequent salinity problems. The following three examples are all from projects concerned with the rehabilitation of irrigation schemes by the installation of drainage.

14.8.2 Indus Plain, Pakistan

A 600 000 hectare area of the Indus alluvial plain in Baluchistan, Pakistan, has been irrigated for many years by a network of canals. Water levels have risen as a consequence, until approximately 40% of the surrounding land is underlain by a water table less than 1.5m below the surface.

In a survey of 61 watercourses in both Baluchistan and Sind, the mean and optimum losses were given as shown below, but the method of calculation is not given.

	Watercourse delivery losses (%)	Field application losses (%)
Optimum	5-9	8-26
Mean	44-45	30-32

A project was commissioned to investigate the drainage potential of this area and the drainable water surplus calculated for sub-regions. The following figures show the drainable surplus estimated in this and earlier reports, in terms of mm/y over the whole area. The figures can be assumed to be of the same order as the recharge and give an indication of the high recharge values attributable to irrigation.

Sub-region		Area (ha)	Drainable surplus (mm/y)
N. Rohri	(1965)	203,000	3934
	(1969)	203,000	4096[1]
W. Nara	(1965)	139,000	1920
	(1980)	189,000	3517[1]

[1] Assuming 51% losses from the watercourses.

Reference:

SIR M. MACDONALD & PARTNERS LTD / HUNTING TECHNICAL SERVICES LTD, 1982. Left Bank outfall drain project preparation (status report). WPDA, Pakistan.

14.8.3 Noubaria Canal, Egypt

The Noubaria Canal in northern Egypt has been established as a water supply since the end of the last century. The recharge to the underlying Nile delta alluvium was indicated by the pronounced ridge in the groundwater contours around the canal, and a steady-state seepage loss has been assumed. More recently, there has been a substantial irrigation development on the west bank of the canal, recharge from which has upset the steady state balance and resulted in the canal now acting as a discharge line for irrigation water recharged to the west.

The pre-development recharge was calculated by both flow net analysis and, more accurately, steady state modelling of the original groundwater mound. The model covered a 4700km² area, largely to the south and west of the canal, with two layers of 199 nodes each. The recharge can be considered as solely resulting from canal seepage and the model was calibrated with a canal recharge of $79 \times 10^6 m^3/y$. Over its 80 km length this gives a recharge of 980 000 $m^3/km/y$ or 2.68 $m^3/m/d$.

The model was then adapted for transient flow and the effects of the irrigation development to the west included. Crop consumptive use was estimated at 20% of the total water supply, and surface runoff thought to vary from 10-30%, with the higher runoff values in the more waterlogged areas. On this basis the initial estimates of recharge from irrigation were 50-70% of the water delivered to the command area; for annual water deliveries of 1500-2000mm this gives a range for potential recharge of 750-1400mm/y.

The annual water balance obtained from the model, using data from 1969 and 1976, is shown below:

Water balance components ($10^6 m^3$)	1969	1976
Underflow	1	1
Potential irrig. rech.	322	654
Total potential inflow	323	655
Underflow	53	60
Seepage to surface	10	63
Rejected recharge	41	330
Total resulting outflow	104	453
Increase in storage	198	202

The figures obtained for rejected recharge are sensitive to changes in the specific yield of the alluvium, so the balance cannot be said to be uniquely correct.

Reference:
FAO / UNDP, 1977. Control of waterlogging and salinity in the areas west of the Noubaria Canal, (Technical report no.1). Ref. no. AG:DP/EGY/73/048, FAO, Rome.

14.8.4 Case 3

This example is from a low lying arid region of 400km², where irrigation water for the fertile alluvium is supplied by a canal network. For the purposes of the model the project area is incorporated into a larger area (2400km²), for which the boundary conditions are more easily defined. The model is bounded to both the south and east by major canals and the general flow direction is from the south to a northern coastal boundary. A distributed two dimensional horizontal model was used, modelling the alluvial aquifer and a 10-20m thick semi-confining layer.

The groundwater inflows and outflows obtained for a steady state simulation are as follows:

Inflow source	Inflow volume (m^3/d)
Canal to south	52 800
Canal to east	1 120
Across the western boundary	10 860
Implied internal recharge	15 600
Total inflow	80 380

Outflow source	Outflow volume (m^3/d)
Canal to south	720
Canal to east	11 350
Across the northern boundary	1 200
Implied internal discharge	67 330
Total outflow	80 600

The model area is recharged by losses from the canal to the south, at the rate of 380 000 m^3/km/y or 1.04 m^3/m/d. The canal to the west acts as a discharge line along most of its length, receiving a net groundwater inflow of 10,230m^3/d.

The implied internal recharge and discharge refers to the gains and losses required to balance the internal modelled and measured head levels and the boundary input and output. The process is of upward and downward flow within the low permeability semi-confining layer, and results in the waterlogging of the discharge areas.

14.8.5 Case 4

The following irrigation scheme, covering an area of 15km^2, is in an arid region of the Middle East with an average annual rainfall of 50-250mm and potential evaporation of > 2000mm/y. Irrigation water is supplied from diverted wadi baseflow. An approximate water balance has been derived for the summer and winter seasons:

Water balance components ($10^6 m^3$)	Winter	Summer
Diverted baseflow	16.6	8.3
Conveyance canal losses	- 1.6	- 0.8
Available irrigation water	15.0	7.5
Evapotranspiration loss	- 7.5-15	- 0.75-1.5
Potential irrig. recharge	<7.5	6.0-6.75

The balance has been derived using the following criteria:

(i) Canal seepage losses are taken to be 10% of the supply, based on earlier field observation.

(ii) Actual evapotranspiration loss is taken as 50-100% of potential loss where potential loss is 100% of the available water. A reasonable AE would be 75% of the PE.

(iii) During the summer the irrigated area is only 10% of the winter area. The evapotranspiration loss is consequently assumed to be a factor of 10 lower, and the remaining water is potential recharge.

A recharge of $9x10^6 m^3/y$ or 600 mm/y was chosen as the initial input to a two dimensional horizontal groundwater model. The first simulation resulted in a marked head loss over the area. As a result of further borehole information the model was altered to represent a two aquifer system and the model was subsequently re-run showing an improvement in the correlation between measured and simulated heads.

15 RECHARGE DUE TO URBANIZATION

15.1 Introduction

Urbanization can radically alter the entire water balance of an area (Fig. 15.1). Water is imported into all large urban areas, increasing all items of the water balance. A micro-climate may develop, particularly in humid zones, with changes in temperature, humidity, wind speed and air clarity. These changes can lead to changes in precipitation, potential and actual evapotranspiration (Hall, 1984).

The increase in impermeable area changes the surface and groundwater hydrology. Infiltration and direct recharge decrease. Surface runoff is increased, although this water may become indirect recharge from a river bed. Where groundwater levels are high, sewers will collect groundwater.

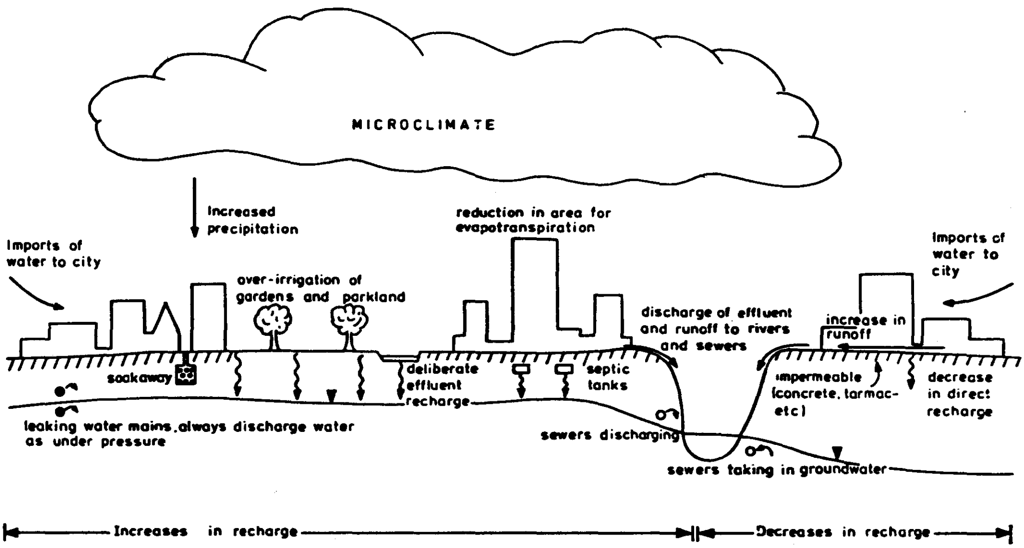


Fig. 15.1 Urban effects on groundwater and surface water hydrology

Urbanization introduces several new sources of recharge:

- septic tanks and leaking sewers
- leaking water mains
- over-irrigation of domestic and municipal gardens
- deliberate recharge of effluent
- deliberate recharge of storm runoff
- unintentional recharge of effluent and storm runoff discharged to rivers.

Table 15.1 gives two examples of arid and semi-arid aquifers which are strongly influenced by these additional sources of recharge. The Doha data are hydrological estimates; the Lima values have been confirmed by groundwater flow modelling. The Lima case is presented in more detail in Section 15.5.1, and a case study of Bermuda (annual precipitation 1460 mm) is given in Section 15.5.2.

Many of the effects of urbanization outlined above are beyond the scope of this manual. The remainder of this chapter will concentrate on water supply; that is water mains, septic tanks and sewers. Only brief comments and reference to other work will be made for the other potential recharge sources.

Table 15.1 Examples of aquifers with increased urban recharge

	Doha Qatar	Lima Peru
Year	1982	1985
Area of aquifer (km^2)	294	400
Total recharge ($10^3 m^3/d$)	92	1155
Recharge from:		
cross boundary flows	ne	171
rivers	ne	410
irrigation losses	ne	196
park irrigation	38	55
leakage of mains	25	323
sewers and septic tanks	18	ne
rainfall	11	0
Total urban recharge (mm/yr)	100	>350

Taken from Lerner (1988). ne = not estimated, or included with another recharge component.

15.1.1 Whether to estimate urban effects on recharge

Sections 15.2-15.4 show that many detailed data can be needed to estimate the effects of urbanisation on recharge. The question naturally arises whether the effort needed justifies the resulting improvement in accuracy of a groundwater model. The answer depends on the accuracy required (a reconnaisance study needs lower accuracy than a design study), on the size of the urban area, on the ratio of water supply to precipitation and other recharges, and on the nature of the city. No firm rules can be laid down, but the following examples may serve as guidelines.

The annual amount of water imported into a city for water supply may range between 300 and 5000 mm, the latter applying to very dense urbanization such as central Hong Kong. If sewage is exported from the city then, as a first estimate, 25% of this imported water will become recharge because it leaks from the pipe network. In humid areas this recharge may balance the loss of precipitation recharge caused by impermeable areas, and the overall effect of urbanisation on recharge will be small. However in arid and semi-arid areas leakage recharge will always be significantly larger than precipitation recharge.

In cities where sewage is not exported, as much as 90% of imported water may recharge groundwater.

Cities which use local groundwater or roof catchments for water supply generally have smaller effects on recharge than those that import water.

The following guidelines may help in deciding whether to quantify the effects of urbanization (they are not designed to estimate recharge):

(i) Urbanization effects should always be quantified if the city covers more than 25% of the aquifer because variations in spatial distribution will become significant.

(ii) The difference between total urban recharge and recharge to the same ground without the city will probably be within 25% under the following conditions:

- rainfall in range 500-1000 mm/yr
- water supply of same order of size as rainfall
- water supply imported from outside city
- about 25% leakage from water mains
- sewerage system removing waste water from city
- storm or combined sewers collecting runoff
- moderate housing density.

(iii) A rough estimate of urban recharge can be obtained by multiplying natural recharge by the following factors:

- rainfall > 1000 mm/yr	:	x 0.75
- rainfall < 500 mm/yr	:	x 2
- rainfall < 250 mm/yr	:	x 4
- all water supplied by groundwater from within city	:	x 0.25
- water supplied by roof catchments	:	x 2
- septic tanks used for waste water	:	x 4
- no storm sewers for runoff	:	x 2
- undersized storm sewers	:	x 2
- high housing density	:	x 2
- low housing density	:	x 0.75
- multiply by (water supply/rainfall)		
- multiply by (%age leakage rate/25)		

(iv) If the net factor from (iii) is more than ±25% then detailed estimates should be considered for modelling purposes.

15.1.2 Groundwater quality in urban areas

Urban areas almost always lower the quality of the groundwater below them and there are certainly high risks of pollution. Potential sources of pollution include:

- septic tanks and leaking sewers
- waste disposal (landfill) sites
- industrial premises, where spillages and leaks of chemicals can be common
- road and rail networks, where accidents and spillages of wastes and chemicals occur
- oil and chemical pipelines.

Therefore, even if recharge is high in urban areas, the resulting resources cannot always be relied upon as a source of potable water. Examples of polluted groundwater in Bermuda, Milan and Birmingham are given by Thomson & Foster (1986), Cavallaro et al. (1986) and Lloyd et al. (1988) respectively; most other cities have polluted their groundwater to some degree.

15.2 Methods of estimating urban recharge

Many of the urban sources of recharge are steady or have random fluctuations unrelated to precipitation. They are often small and are usually at unknown locations. These factors prevent the use of most of the methods outlined in Chapter 10. Water balance techniques are the most useful.

It is possible in principle to use tracers to estimate various components of urban recharge. For example, imported water which is put into supply may have different concentrations of solutes from the local groundwater. Mass balances of solute based on averages of widespread sampling may indicate proportions of imported water. However there are no reported case histories of such a method and the potential errors are high. For example, leakage from water mains is often localised, and its effects may be missed in a sampling programme. It may only mix in the top part of the groundwater body, invalidating sampling in open holes where the water column is mixed. In all cities water supply and so urban recharge is growing; samples will neither represent the current position nor a steady state, but an average over an unknown number of years. However tracers have successfully been used to identify sources of urban recharge (Lerner, 1986).

15.3 Water supply and sewage

There are a great many components of the balance of water supply and sewage in an urban area (Fig. 15.2). Recharge is only a small component, made up as follows:

$$\begin{aligned} \text{recharge} = \; & \text{leakage from water mains} \\ & + \text{external losses from consumers' properties} \\ & + \text{leakage from sewers} \\ & + \text{flow to septic tanks} \\ & + \text{deep percolation from domestic irrigation} \\ & + \text{deep percolation from municipal irrigation} \end{aligned} \tag{15.1}$$

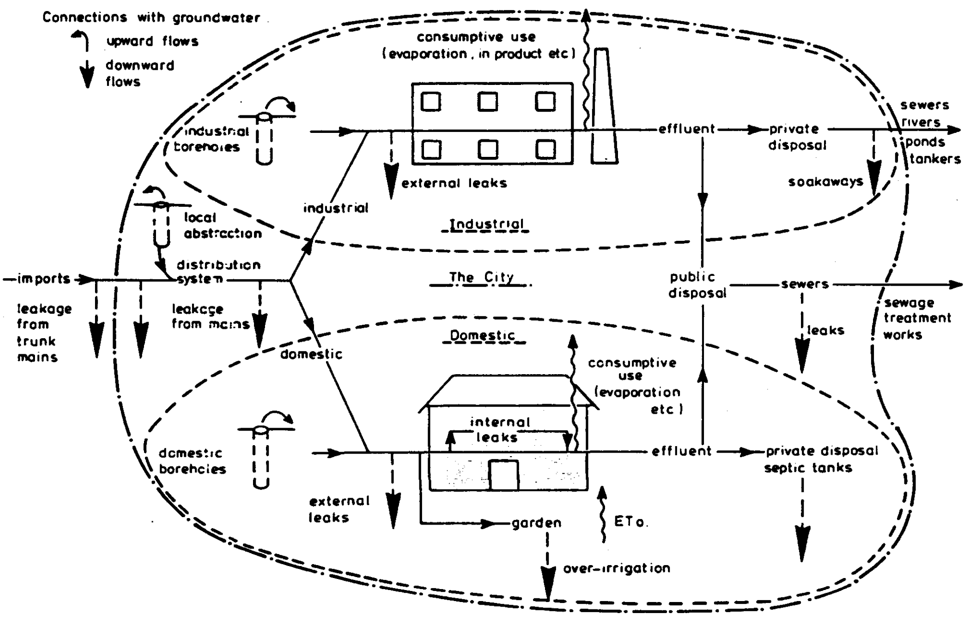


Fig. 15.2 Flows of water supply and sewage in a city

Considering all these components individually, with their associated errors, can lead to a large accumulated error in the recharge estimate. To reduce errors, it is preferable to consider the overall balance and estimate net recharge as follows:

$$\begin{aligned} \text{net recharge} = \; & \text{imports of water} \\ & + \text{local abstractions of groundwater} \\ & - \text{consumptive use} \\ & - \text{effluent leaving area} \end{aligned} \tag{15.2}$$

Different approaches are needed for small, medium and large communities. Sections 15.3.1-2 discuss small and medium communities respectively. Section 15.3.3 discusses the components of water supply and sewage which will need to be considered for large communities and in some other cases.

15.3.1 Small communities

In arid and semi-arid areas, individual farms and villages usually rely on local groundwater supplies. Waste water is disposed of to the ground surface or to septic tanks. Thus most of the groundwater taken out is recharged locally. For modelling purposes, the net effect is of a small withdrawal equal to the consumptive use:

$$\text{net withdrawal} = \text{population} \times \text{consumptive use per person} \qquad 15.3$$

In the absence of local information, consumptive use can be assumed to be 20 l/head/d (loosely based on Tables 5.1 and 5.2, Dangerfield, 1983). The quantities will normally be too small to justify more detailed study.

15.3.2 Medium sized communities

Larger communities import water, either from more distant groundwater sources or from perennial surface water sources. The net effect is normally an increase in groundwater quantities in the urban area (matched by a decrease in resources elsewhere).

If the waste waters are recharged locally, for example through septic tanks, then net recharge can be estimated as:

$$\text{net recharge} = \text{imports of water} - \text{consumptive use} \qquad 15.4$$

This assumes that imports of water are measured and that consumptive use can be estimated (see later). Considering the components of recharge (eqn 15.2) is rarely likely to improve the accuracy because of the need for complete and accurate data on all components.

Table 15.2 Daily consumption of water in various cities (l/head/d)

City	Date	Total	Domestic	Reference
Cairo	1958	180		vdL
Sydney	1970	446		vdL
Mexico City	1969	323	227	vdL
Baghdad	1969	172		vdL
Jerusalem	1969	165		vdL
Jerusalem	1979	290	210	ECR
Houston	1970	950		vdL
Amman	1969	64		vdL
Haifa	1979	101	69	ECR
Barcelona	1979	267	190	ECR

vdL = van der Leeden, 1975. ECR = Reed, 1980.

The amount of water imported clearly depends on the population and the amount of local supplies. It will also depend on the the way water is distributed; consumption is higher when water is piped to individual homes than when it is delivered by tanker or to communal standpipes. Standard of living and the condition of water mains affects total consumption. It is not possible to provide "typical" values that could be applied in the absence of local data, as the range of consumption values in Table 15.2 shows. Where imports are not measured, it will be necessary to consider the components outlined in eqn 15.1.

Consumptive use of potable water in a city (that is water evaporated or put into industrial products), has three main components; personal domestic use, evapotranspiration if potable water is used for irrigation, and industrial use. Personal domestic use is about 20 l/head/d. If either of the other two components is significant, the approach of eqn 15.4 may be too inaccurate, and all the components of eqn 15.1 must be considered, as discussed in the following paragraphs.

15.3.3 Components of water supply and sewage recharge

Leakage from water mains cannot be measured directly. There are two common practices for estimating leakage. Firstly it can taken as a proportion, L, of the water supplied:

$$\text{leakage} = L \times \text{supply} \qquad 15.5$$

Typical values of L are 25-35% but values of 10-60% have been reported. Secondly, leakage can be estimated as the difference between supply (input to the system) and use (output from the system):

$$\text{leakage} = \text{supply} - \text{use} \qquad 15.6$$

Eqn 15.5 is easier to apply but has higher potential errors. Its accuracy depends on the accuracy of L, which is either guessed, transposed from another city, or estimated for a small area using eqn 15.6. Therefore this section will concentrate on the use of eqn 15.6.

The way in which eqn 15.6 is used will depend on how water use is measured in a city, for example whether domestic consumers are metered, and on the time period chosen. Over periods of a day, month or year, eqn 15.6 can be written as:

$$\begin{aligned} \text{leakage} = \ & \text{total supply} \qquad 15.7 \\ & - \text{bulk transfers out of area} \\ & - U_1 \times \text{metered industrial and domestic use} \\ & - \text{unmetered industrial and domestic use} \\ & - \text{municipal use} \end{aligned}$$

where U_1 is a factor to allow for the commonly occurring under-recording of meters which ranges from 1-10% (Dangerfield, 1983). Unmetered use is frequently estimated from population and per capita consumption. The latter varies greatly between cities (Table 15.2) and local, or at least national, information is essential. In this context, municipal use includes street cleaning, firefighting and sometimes irrigation of parks and roadside verges. Phillips (1983) gave estimates of 2 l/head/d for municipal use in the UK, with some of the principal uses being street cleaning (0.25), firefighting (0.06) and building (0.83 l/head/d).

It is a common assumption that leakage is roughly constant throughout the day. Therefore more accurate estimates of leakage are obtained by applying eqn 15.6 at times when use is small, and when leakage is a large proportion of supply. This is typically in the early hours of the morning, and corresponds with a common technique in the water supply industry of 'minimum night flow' (MNF) measurements. MNFs are measured by adjusting valving so that all flow to a district is through one point where it is measured. The lowest flow rate is then recorded, usually at 02:00-04:00. Ideally, all meters are read at the same time to determine use. In this case, eqn 15.6 is written as:

$$\begin{aligned} \text{leakage} = \ & \text{MNF} \\ & - \text{bulk transfers out of area} \\ & - \text{refilling of reservoirs and} \\ & \quad\ \text{domestic water tanks} \\ & - U_2 \times \text{metered use} \\ & - \text{unmetered use} \end{aligned} \qquad 15.8$$

Municipal use is negligible at these times of day. A different factor, U_2, for under-recording of meters applies at night because meters behave differently at low flow rates.

A common approach to unmetered domestic usage in the UK is to adopt a MNF value of 2 l/property/h. However Philip (1985) reported values from individual houses of 0.5-9.0 l/property/h. If meters are not read during the MNF exercise, metered night use must be taken as a proportion of average metered supply. Philip (1985) showed for the UK that industrial and commercial night use was typically 50% of average rates, a surprisingly high value. If there is no information on night use, take leakage to be 50% of MNF (Lerner, 1986).

<u>External losses on consumers' properties</u> make up the second component in eqn 15.1. The nature and accounting of leaks on the consumers' side of their meter or stop valve varies between cities and local knowledge is essential. In some cities, leaking water flows to the sewers and can be included in potential sewer recharge. In other cases, a good proportion of the leaks are within or onto the ground and can be considered as recharge.

Estimates of the leakage from water mains, as produced by water supply authorities, may not take into account leaks from the service pipe connecting the water main to the house, and these are often the leakiest parts of the system. This will be reflected in high values of legitimate night use per property, which includes this type of leakage as a component. It would then be prudent to assume a lower value of night use within each property, of perhaps 2 l/property/h.

Septic tanks can be assumed to recharge all the water they receive. If they are the only means of waste water disposal, then water supply recharge is best estimated as (total supply - consumptive use) to avoid discovering the components.

Sewers. All the published quantitative work on the interaction between sewers and groundwater is about infiltration into the sewers (eg Pluhowski & Spinello, 1978, Steketee & Heinecke, 1984). This is a major variable when designing sewer networks and treatment works (Escritt, 1984). Infiltration into sewers below the water table can be high; Hannigan (1984) reports a 450mm diameter sewer collecting $9m^3/d/m$. It is reasonable to assume that high leakage rates are possible for sewers above the water table. In a study of the Triassic sandstone aquifer under Liverpool in the UK, storm sewer leakage was about equal to water main leakage (University of Birmingham, 1984). A number of groundwater quality studies have shown that sewers do leak, but have not estimated the flows involved (Hornef, 1985; Seyfried, 1984; Duyvenbooden & Loch, 1983).

Estimating the amount of sewer leakage is difficult. Details of sewer construction, their "leakiness" and the properties of the adjacent soils will be unknown. Inflows and outflows to sewers are rarely measured, and leakage is likely to be within the measurement error of flows. One view is that leakage may be low because sewers are generally unpressurized (Lerner, 1986) but there is as yet no firm evidence for this. The difficulties of estimating sewer leakage emphasise the value of taking an overall approach to a city as outlined above.

Deep percolation from over-irrigation is briefly discussed in Section 15.4.5 below.

Roof catchments are widely used on islands to capture a high proportion of rainfall and conserve water resources. If septic tanks are used to dispose of waste water, the use of roof catchments will increase recharge because the consumptive use of water will be lower than the evapotranspiration that would have occurred. An example from Bermuda (Thomson & Foster, 1986) is summarised in Section 15.5.2.

15.4 Other urban recharges

15.4.1 Direct recharge

Urbanisation increases the impermeable area. If the storm runoff from this impermeable catchment is channelled to storm sewers and away, then direct recharge will be reduced by urbanisation. However there are circumstances when recharge is not reduced but increased. Some Dutch work (van Dam & van de Ven, 1984) has shown that surfaces covered with bricks and tiles (such as pavements and car parks) are remarkably permeable. In addition the surface covering reduces evaporation.

If storms are rare there may be no storm sewer system; runoff will infiltrate along the edges of the impermeable areas, or along nearby drainage channels and depressions. Thus total recharge from rainfall may be the same or higher than if the city was not present.

Soakaways are often used to dispose of storm runoff. They may be for domestic properties only, for road drainage, or for general drainage. In all cases they will increase the amount of direct recharge.

If direct recharge is potentially significant, careful studies of storm runoff will be needed to quantify the effect of the city; these are beyond the scope of this manual, but are dealt with in standard texts such as Hall (1984). From the point of view of a model with nodal areas of 1 km^2 or more, the exact location of recharge is immaterial. In this case, precise details of runoff routes are not needed, and an overall balance of storm flows may be sufficient.

15.4.2 Increased surface water flows

Flows in rivers may be increased by both storm runoff and effluent discharges, both of which may cause increased recharge through the river bed. In principle such recharges can be estimated by the methods discussed in Chapter 12.

15.4.3 Recharge basins for storm runoff

In some urban areas storm runoff is channelled to retention or recharge basins. The purpose may be to delay or dispose of runoff to prevent flooding downstream, or to increase groundwater resources. As an example, recharge basins on Long Island, New York are described by Seaburn & Aronson (1974). Storm water management techniques are beyond the scope of this manual. The reader can find methods for estimating or measuring storm runoff in standard texts (e.g. Hall, 1984) which can be used in water balance approaches to recharge estimation. Methods using Darcy's law are discussed in Chapter 12.

15.4.4 Effluent recharge

The city's effluent may be disposed of by recharging through special lagoons, discharging to a river bed, spraying or spreading, or exceptionally, through boreholes. Such deliberate artificial recharge is beyond the scope of this manual. Standard texts include Todd (1980) and Dept. Econ. and Social Affairs (1975). Many of the methods outlined in Chapter 10, including water balances, Darcian methods and tracers, may be applicable.

15.4.5 Over-irrigation

Parks and gardens are irrigated for aesthetic rather than commercial reasons. The irrigation is often done by unskilled people. The amount of water applied rarely depends on plant water needs, but on the affluence of consumers, pricing policies for water supplies and, in the case of municipal parks, on bureaucratic procedures. For these reasons over-irrigation is normal with many multiples of the potential evapotranspiration being applied. Excess water percolates deep to recharge groundwater. Quantities can, in principle, be estimated by any of the methods described in Chapter 14.

15.5 Case studies

15.5.1 Lima, Peru

Lima, the capital of Peru, stands on the combined alluvial fans of the Rimac and Chillon rivers. The aquifer covers about 400 km^2 and is up to 500m thick. Over half of the surface is now urbanised, with the remainder used for irrigated agriculture.

Rainfall is too low to provide recharge and the groundwater system is entirely replenished by river water by a variety of routes including:

- direct from the river
- from river to canal to irrigated land, with excess irrigation becoming recharge
- from river and boreholes to the potable supply network, with substantial leakage losses becoming recharge
- from river, via canal or tanker, to parks for irrigation, with excess becoming recharge.

Water balance methods were used to estimate the various recharges; these estimates were further refined by calibration of a groundwater flow model. Leakage from water mains is of most interest at present and was estimated as follows. Minimum night production of water for the whole city was known for a few dates and was about 69% of average production. Gross leakage was estimated by deducting estimates of:

(i) legitimate night-time consumer consumption, for which no local data was available and so a range of 2-13% of average supply was adopted;

(ii) bulk transfers out of the area, measured as 2% of average supply;

(iii) refilling of service reservoirs at night, found by measurement of a few and extrapolation to be 5-10% of average supply.

Only part of the gross leakage becomes recharge because some leaks within consumers' premises and flows to the sewers. The first estimate of recharge from leakage was 40% of average supply, but this has been revised downwards in the modelling studies to 30%. As can be seen in Table 15.1, recharge from leakage is still the second largest recharge to the aquifer, providing about 28% of the total.

Further details of the Lima study can be found in Lerner et al. (1982), Lerner (1986) and Wild & Ruiz (1987).

15.5.2 Bermuda

This summary of the effects of urbanisation on groundwater recharge in Bermuda, a limestone island in the Caribbean, is based on a paper by Thomson & Foster (1986).

The islands of Bermuda are formed from a sequence of Pleistocene aeolian limestones. There are five known fresh groundwater lenses, of which the largest is the central lens (Fig. 15.3a). It has a fresh water nucleus up to 10m thick which covers 630 ha, and a substantial brackish fringe. The 1-99% salinity transition zone is 5-6m thick under much of the lens, but thicker near the coasts (Fig. 15.3b).

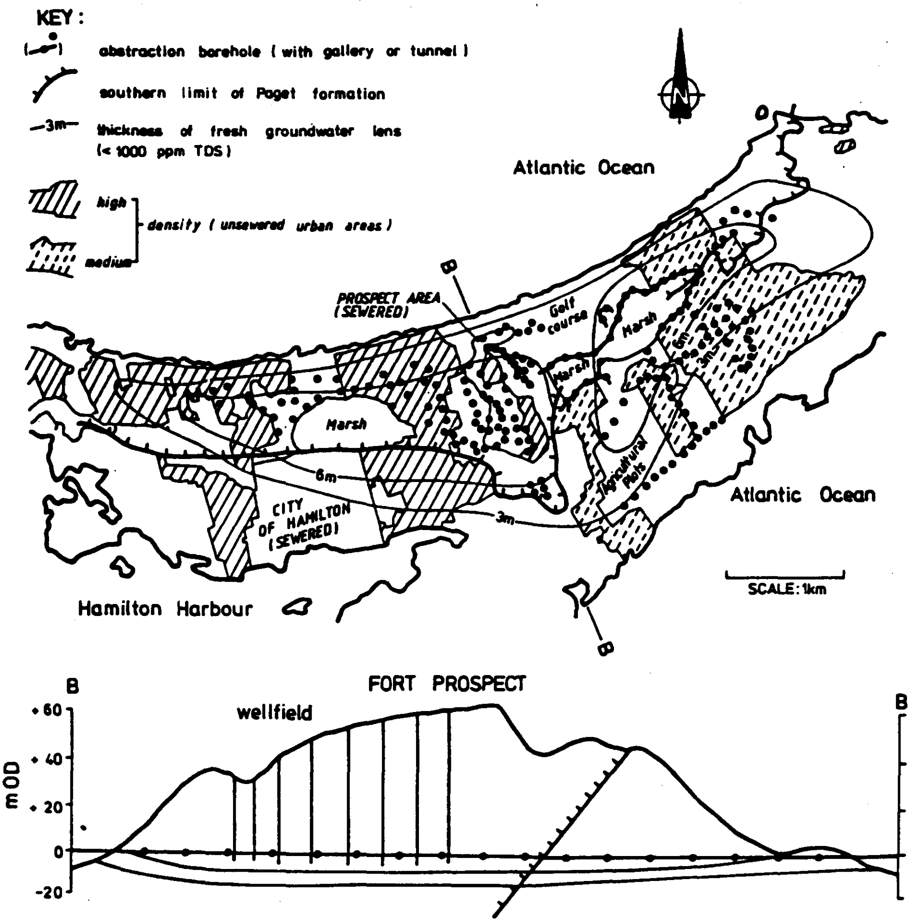
Fresh water has always been regarded as a valuable commodity in Bermuda. Rainwater is collected from roof catchments (covering a prescibed 80% of roof area) and stored in large underground tanks. With the high rainfall of 1460 mm/yr, this system provides the primary water supply for almost all households. Only 8% of households are sewered, the remainder discharging to soakaways.

More than 150 boreholes are used to abstract 4500 m^3/d from the central groundwater lens which is supplied (after treatment) to municipal and commercial establishments.

The calculation of recharge is summarized in Fig. 15.3c. The starting point is naturally vegetated grass and woodland, from which a soil moisture model estimates 75% of average rainfall is evaporated, leaving a recharge of 365 mm/yr. The small agricultural area is thought to have higher recharge of 440 mm/yr because of the sparser and occasionally absent cover, and less well developed root system.

Recharge in the urban area depends upon housing density, which will change the ratio of roof+road to vegetated areas. Three different densities have been identified. The vegetated parts are assumed to have the same recharge rate as naturally vegetated areas (365 mm/yr). Rainfall on roofs and roads either passes direct to drainage soakaways, or via domestic water supply systems to waste water soakaways; in both cases a loss of 10% has been assumed. On this basis, annual recharge in the urban area varies between 410 and 740 mm, that is up to twice natural recharge.

The paper by Thomson & Foster (1986) gives information on the existing and potential pollution of Bermuda's groundwater. In summary, there are already significant levels of nitrate and some microbiological pollution. A definite risk of pollution by synthetic organic chemicals has been identified.



precipitation 1460mm/a

land use	RESIDENTIAL			GRASS & WOODLAND	AGR	MARSH
residential density (persons/ha	HIGH 65/ha	MEDIUM 35/ha	LOW 5/ha	0/ha		
area/population	150ha 9750	100ha 3500	140ha 700	150ha	25ha	65 ha
ratio roof + road to vegetated area	40:60	20:80	5:95	0:100		
evotranspiration	75% on vegetated part / 10% on roads + roofs				70%	100%
annual recharge	740mm	555mm	410 mm	365mm	440 mm	none
recharge volumes	1110 Ml/a	555 Ml/a	575 Ml/a	550 Ml/a	110 Ml/a	

lens area 690ha
population 13,950

of rainfall

2900 Ml/a
(7950 m^3/d)

Fig. 15.3 Bermuda central groundwater lens.
(a) public water supply boreholes and landuse
(b) cross-section with 1% and 99% salinity contours on the fresh/saline interface (c) estimation of recharge

16 NET RECHARGE OVER A REGION

16.1 Introduction

The previous chapters have dealt with the various components of recharge to an aquifer, based on the origin of the water (precipitation, irrigation, etc) and its route to groundwater. Alternative approaches are to estimate the total recharge, or the recharge net of some of the outflows which are less easy to calculate.

Such approaches seek to avoid describing any of the complex processes controlling recharge and groundwater movement, such as evaporative losses from the unsaturated zone, rivers with both influent and effluent reaches, and phreatophyte evapotranspiration.

There are four types of methods described in this chapter:

(i) Storage change or water table rise, ie

$$\text{total recharge} = \text{change in saturated volume} \times \text{specific yield} + \text{outflows} \qquad 16.1$$

(ii) Discharge from the aquifer, that is assuming that the measured discharge of groundwater across the boundary of a study area is equal to the net recharge inside the study area.

(iii) Inverse techniques, that is inverting the groundwater flow equation to solve for recharge instead of groundwater heads.

(iv) Tracer techniques, either estimating residence time of water in an aquifer, or modelling the distribution of several tracers throughout an aquifer.

All the methods have advantages and disadvantages as discussed below. Numbers (i), (ii) and (iv) are useful as checks on a water balance for the aquifer calculated by considering all the components of recharge, by the methods outlined elsewhere in this manual. The methods are only accurate for groundwater systems which are in hydraulic equilibrium. A summary of the methods outlined in this chapter is given in Table 16.1.

16.2 Water table rise

The total recharge to an aquifer is often estimated from the volume of water stored as the water table rises during the wet season. The method clearly only applies to aquifers with a well defined recharge season. Allowances are made for pumping wells and other discharges as required, and recharge is estimated as:

$$r = (\delta s + \Sigma Q_A \ \delta t + V_D) \ / \ A \ \delta t \qquad 16.2$$

Table 16.1 Summary of methods of estimating net recharge over a region

This table is not a substitute for the longer discussion of the methods given in the main text of this chapter.

Method	Water table rise	Discharge from basin	Inverse methods	Transit time models	Mixing cell models
Section	16.2	16.3	16.4	16.5.1	16.5.2-16.5.3
Applicability	Unconfined, unstratified, underexploited aquifer with distinct recharge seasons or events	Groundwater systems with well defined boundaries and outflow points	Simple groundwater systems in steady state, preferably with distinct recharge areas	Small groundwater systems with single recharge source	Small or underexploited groundwater systems
Accuracy	Poor (when assumptions violated) to reasonable	As good as knowledge of system and accuracy of outflow measurements	?	Poor to reasonable	?
Data requirements	Hydrogeological conceptual model. Specific yield. Pumping and other outflows. Groundwater hydrographs	Measured flows at all outflows, including boreholes. Hydrographs and specific yield if not steady state	Conceptual model. Mathematical model. Spatial pattern of permeability. Detailed groundwater head maps	input and output concentrations of tracer over flushing time of aquifer	Conceptual model. Input and spatial pattern of tracer concentration(s)
Ease of use	Straightforward	Straightforward, if outflow data can be obtained	Difficult	Reasonably simple	Moderate to difficult. Computer skills needed
Type of estimate	Areal average for period of hydrograph	Long term average over whole aquifer	Long term average at each node	Long term areal average	Long term average at each node
Costs	Low, if observation wells exist	Low if data already exist, otherwise moderate to high	High - specialised skills needed	Low if data already collected	Moderate
Time required	Analysis is short, data collection can be long	Analysis is short, data collection can be long	Moderate, say 3-12 man-months if data already collected	Short if data already collected, long if input/output concentrations must be measured	Moderate: model must be developed and calibrated. Longer if data must be collected from field

where r : total recharge per unit area (L)
δs : volume of water stored between lowest and highest water table positions (L^3)
Q_A : abstraction rate from wells pumping during the water table rise (L^3/T)
δt : the time interval between high and low water table positions (T)
A : area of the aquifer (L^2)
V_D : the volume of water discharged to springs, river bed seeps, etc (L^3)

The volume of water stored is calculated by (i) mapping and contouring water level rises, (ii) measuring areas between contours, (iii) summing the aquifer volume within which the variation in water level takes place and (iv) converting to volume of water using the aquifer's specific yield, as follows:

$$\delta s = S_y \; \Sigma_i \; 0.5 \; (c_i + c_{i+1}) \; A_i \qquad 16.3$$

where S_y : specific yield
c_i : value of contour i of water table rise (L)
A_c : area between contours c_i and c_{i+1} (L^2)

The method is appealing because of its simplicity and because it appears to be a direct measurement of total recharge. There are conceptual difficulties with the meaning of "specific yield" and "water table" that can cause the method to give unrealistic results, and a number of problems of practical application that need attention. These are discussed below; most of the points can be referred to Fig. 16.1.

Specific yield is defined as the amount of water per unit area released from storage for a unit fall in water table level. However, lower permeability materials show time-dependent effects in drainage and imbibation; time to equilibrium may reach years for some clays, a common constituent of alluvial aquifers. Hysteresis, with different wetting and drying properties are common (Fig. 11.5). During wetting, residual air will be left until it dissolves.

Figure 16.2 shows schematically how changes in lithology can cause unexpected and variable specific yields. Such lithological variations are common in alluvial aquifers. Meaningful estimates of specific yield are difficult to obtain, with a range of 0.02-0.40 possible by different methods for the same alluvial material (Rushton, Chapter 23). In general, a "drainable porosity" measured in the laboratory over a reasonable number of samples is the appropriate value to use. Estimates from pumping tests are often noticably different because they are derived for short times. It is common in groundwater flow modelling, which is concerned with medium to long-term changes, to find that specific yields from pumping tests do not work. If no laboratory measurements are available, it is better to use the "standard" values which are often found in textbooks, as for example in Tables 16.2 and 16.3.

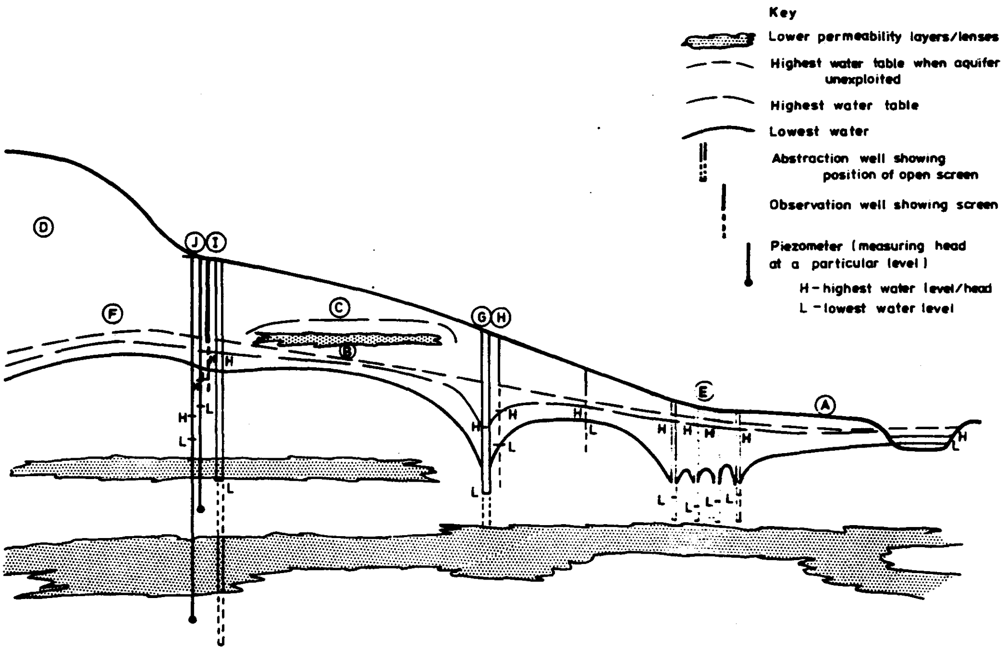


Fig. 16.1 Water table rise method of estimating total recharge. See text for explanation

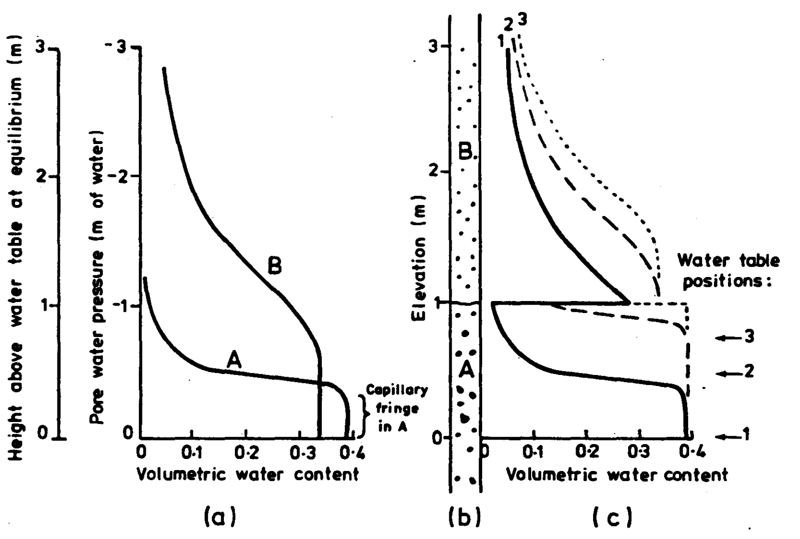


Fig. 16.2 Moisture content and specific yield.
(a) schematic equilibrium moisture-pressure curves for : A-coarse sand, B-well graded fine sand
(b) example lithology
(c) moisture contents for three positions of water table. Specific yield between positions 1 and 2 = 57%, between 2 and 3 = 28%

Table 16.2 Average values of specific yield. Taken from Todd, 1980

Material	Grain size	Specific yield (%)
Gravel	coarse	23
	medium	24
	fine	25
Sand	coarse	27
	medium	28
	fine	23
Dune sand		38
Silt		8
Clay		3
Loess		18
Sandstone	medium	21
	fine	27
Siltstone		12
Limestone		14

Table 16.3 Values of specific yield in the zone of water table fluctuation as used in recharge calculations in India. After Sinha & Sharma, 1988

Material	Range of specific yield (%)
Sandy alluvium	12-18
Valley fills	10-14
Silt/clay rich alluvium	5-12
Sandstone	1-8
Limestone	3
Highly karstified limestone	7
Granite	2-4
Basalt	1-3
Laterite	2-4
Weathered phyllites, shales, schists and associated rocks	1-3

<u>The water table</u> is defined as the position where the porewater pressure is equal to atmospheric. A common simplification is often made that the aquifer is fully saturated below the water table, and unsaturated above. This is of course not true, with a saturated capillary fringe and positive moisture contents above the water table depending on a negative pressure (Fig. 16.2a). For shallow water tables (eg A in Fig. 16.1) only a small amount of water can go into storage before the water table reaches the surface because the ground already contains moisture. Gillham (1984) discusses near surface effects of capillary fringes. This effect also occurs when the water table rises through a change in lithology to a finer pored material with greater moisture retention (B in Fig. 16.1; Fig. 16.2b).

The water table reaches its peak, not when recharge has stopped, but when outflows and lateral movements of groundwater exceed the rate at which water is arriving at the water table. Two particular cases are illustrated at C and D. Perched water tables (C) may form on low permeability layers or lenses above the main water table. These will impede the flow of recharge and may spread it well beyond the time of the peak water table. A deep unsaturated zone (D) will cause a similar effect, delaying and spreading out the recharge. Recharge rates may fall as a new growing season starts, or as rainfall declines at the end of the season. Both may cause the water table to decline before recharge actually stops.

Water table rise methods are commonly used in heavily exploited aquifers, as only these have sufficient observation wells. Problems arising from heavy pumping, which is often seasonal to provide irrigation water in the dry season, may include:

(i) major abstractions (e.g. E in Fig. 16.2) may induce inflow from perennial water courses (A) which will reverse during the recharge season,

(ii) the groundwater catchment boundary (F) may alter during the recovery of the water table,

(iii) many observation wells will be pumped in the dry season (E) and part of their water level recovery will be due to the seasonal stopping of abstraction and the subsequent redistribution of water levels in the aquifer,

(iv) wells may be pumped throughout the year at varying rates (G),

(v) unpumped observation wells (H) may be strongly influenced by nearby pumped wells,

(vi) deep boreholes (I) may draw from confined or semi-confined layers at depth. Piezometers, or wells open in these layers, will show much larger drawdowns (J) than the water table. Long after pumping has stopped, these piezometers will show significant drawdowns and there will be vertical flow from the water table aquifer to the deeper layers. This is particularly applicable to alluvial deposits.

Contouring of groundwater is usually subjective because there is rarely enough data to define all the features. For this reason, repeat contourings of an area should be carried out by the same hydrogeologist to ensure that the same biases appear in both maps.

Objective and repeatable methods of contouring and estimating volumes exist. They include computer contouring packages and geostatistical techniques such as kriging (Clarke, 1979; Gambolati & Volpi, 1979; Lovell et al., 1972). Two disadvantages with such methods are their complexity and their reliance on computers. More importantly, they cannot use the subjective information (eg position of pumping wells, faults, rivers) that an experienced hydrogeologist would take into account in hand contouring. Such methods are probably only appropriate for straightforward aquifers with many well-distributed data points.

The discussion above on the possible errors in the water table rise method may seem overlong and pessimistic. It has been included because the method is in widespread use, and its apparent simplicity conceals many simple conceptual models of groundwater processes. The most important of these are (i) a single aquifer with no vertical flows, (ii) specific yield.

Another method of estimating total recharge is to calibrate a distributed groundwater flow model, matching predicted changes in groundwater levels against measured. This can be seen as an alternative version of the water table response method discussed above.

16.3 Discharge from the basin

For an unexploited aquifer,

$$\text{average discharge} = \text{average net recharge} \qquad 16.4$$

For an exploited aquifer, which in general will not be in steady state in an arid/semi-arid environment,

$$\underset{\text{discharge}}{\text{average}} = \underset{\text{recharge}}{\text{average net}} + \underset{\text{depletion}}{\text{rate of storage}} \qquad 16.5$$

Thus an estimate of discharge over the aquifer boundaries provides an estimate of net recharge. The possible routes of discharge from a groundwater basin include springs, rivers, lakes and seas, sabkas, evapotranspiration, abstraction by man and storage changes. These are discussed in the following paragraphs.

<u>Springs</u> are discrete discharge points, usually at changes in lithology. Flows can often be measured. Water use by vegetation on the banks of the spring and its watercourse is often significant; gaugings downstream may require correction for such channel losses.

<u>Rivers</u> receive groundwater either at discrete points or as diffuse seepage. Surface water flows must be separated from groundwater outflows by hydrograph analysis. River-aquifer interactions can be complex, and hydrograph separation made difficult by bank storage and subsurface flows in the riparian zone. Methods of separation are discussed by Todd (1980), Wright, (1980), Chow (1964) and other hydrological text books.

<u>Lakes and seas</u> also receive groundwater at discrete points and as diffuse seepage. Measurements will be possible only in exceptional circumstances. Lee (1977) describes a method of measuring diffuse seepage through lake beds. Otherwise, arcian estimates based on hydraulic head and permeability measurements are most likely to provide satisfactory data.

Sabkas or salt pans are discharge areas for groundwater systems where all discharge is evaporated. Often there is no water visible at the surface with the unsaturated upward flow of groundwater evaporated at or below the surface. Direct measurement of evaporating fluxes is possible in principle but is unlikely to be practicable (Section 11.4.2). Darcian estimates of groundwater flow either in the aquifer towards the sabka or vertically upwards in the evaporation area are possible.

Inverted chloride profiling may provide estimates of upward flow; normal chloride profiling is discussed in Section 11.6.2. Stable isotopes (^{18}O and ^{2}H) are enriched at the evaporating front and in principle can be used to estimate evaporation rates. There are theoretical problems of ensuring a steady state exists, and has existed for sufficient time to develop a characteristic profile, and that isotope profiles are not distorted by occasional precipitation. There are many practical difficulties, including (a) sampling without contamination, heating, or evaporative losses; (b) extracting enough moisture from a zone which has a very low moisture content, and of measuring isotope ratios in local rainfall and water vapour. The theory of the method is given by Barnes & Allison (1983), practical examples by Allison & Barnes (1985), Sonntag et al. (1985) and Yousfi et al. (1985). Allison et al. (1985) concluded:

> "^{18}O and ^{2}H profiles appear to be promising tools for quantitative estimation of groundwater discharge .... however some further work needs to be done to develop the technique fully."

Evapotranspiration by vegetation may be a significant discharge of groundwater in arid and semi-arid areas. This may be shallow rooted vegetation drawing water through the unsaturated zone or phreatophytes tapping the water table directly. The former must be estimated in the same ways as evaporation in sabkas; it may be necessary to allow for seasonal changes or cropping. Estimating phreatophyte evapotranspiration is notoriously difficult, as it varies by several orders of magnitude between species, and also varies with the maturity of the plant, groundwater salinity and the depth to the water table. Some examples of evapotranspiration rates are given in Table 16.4, and an illustration of the relative sizes of evaporation and phreatophyte transpiration is given in Table 16.5.

FAO irrigation and drainage paper 37 (1981) lists five different methods of estimating phreatophyte evapotranspiration of which the transpiration well method is prefered. Maximum and minimum water levels are measured in piezometers both within a phreatophyte stand and outside its influence and the variation in levels is related to phreatophyte uptake and specific yield.

Abstractions by man are through wells, boreholes or drainage. A proportion of the water taken for irrigation or domestic use may well return to the aquifer locally.

Storage changes would not normally be considered as a discharge. However, in arid and semi-arid areas where recharge is low or erratic, springs and other discharges may be sustained by slowly declining water levels in between sporadic bursts of recharge. These may be as frequent as once in two years, or as rare as once in 1000 years (Edmunds & Wright, 1979). In this circumstance, measured discharges will overestimate recharge and allowances must be made for the rate of loss of storage.

Table 16.4 Examples of groundwater use by phreatophytes. After Robinson, 1958

Scientific (and common) name	Depth to water table (m)	Annual water use (mm)
Alnus (alder)	?	1600
Baccharis glutinosa	0.6	3100
(batamote)	1.9	1400
Cynodon dactylon	0.6	870
(bermuda grass)	0.9	720
Juncus balticus (wiregrass)	ns	2400
Prosopis velutina (mesquite, velvet)	?	1000
Sporobolus airoides	0.6	1220
(sacaton, alkali)	1.2	1050

Table 16.5 A comparison of water uptake by phreatophyte transpiration and evaporation. After Hughes & McDonald, 1966

Method of water loss	Depth to water table (m)	Annual water loss (mm)
Evaporation from class A pan	-	2515
Evaporation from bare soil tank	0.9	176
Transpiration by Pluchea servicea (arrowweed)	1.7	2716
Transpiration by Atriplex (saltbush)	1.6	730

16.4 Inverse techniques

The flow of groundwater can be represented by an equation that includes (i) groundwater heads, (ii) aquifer properties, (iii) recharges and boundary fluxes. Knowing the boundary conditions and two of these sets of data allows solution of the equation for the third. Conventional groundwater models solve for groundwater heads, inverse models solve for aquifer properties or, more unusually, recharge.

For steady state flow, the equation of flow is:

$$r = \frac{\delta}{\delta x}\left[T\frac{\delta h}{\delta x}\right] + \frac{\delta}{\delta y}\left[T\frac{\delta h}{\delta y}\right] \qquad 16.6$$

where
- x : horizontal co-ordinate (L)
- y : horizontal co-ordinate (L)
- h : hydraulic head (L)
- T : transmissivity (L^2/T)
- r : recharge (L/T)

Using a finite difference or finite element discretisation, and including the boundary conditions leads to (in matrix notation):

$$\mathbf{A\ h = r_u + r_k} \qquad 16.7a$$

A direct solution for the unknown flows is:

$$\mathbf{r_u = A\ h - r_k} \qquad 16.7b$$

where
- $\mathbf{A}$: coefficients involving transmissivity and constant head boundary conditions (n by n matrix)
- $\mathbf{h}$: groundwater heads (n)
- $\mathbf{r_u}$: unknown recharges (n)
- $\mathbf{r_k}$: known parts of recharge or discharge (n)

although less obvious solution techniques will give more stable and useful results as discussed below.

Inverse techniques to find aquifer properties were a popular research field in the late 1970's. Three problems occur with most methods; great sensitivity to errors in water level and recharge data, low sensitivity in unstressed areas of an aquifer, and non-uniqueness of solutions. These problems are well discussed by Cooley & Sinclair (1976), Neuman (1980), Yakowitz & Duckstein (1980) and McElwee (1982) and have meant that inverse techniques have had little practical application.

Few authors have attempted inverse techniques to estimate recharge. In general they can be expected to suffer from the same problems identified for aquifer property solutions above. However, Smith & Wikramaratna (1981) and Wikramaratna & Reeve (1984) point out that constraining where recharge can occur in a model removes much of the uncertainty and leads to a much improved solution. In arid and semi-arid areas, recharge may only occur in certain areas, for example as mountain front or wadi recharge (Fig. 16.3). Smith & Wikramaratna (1981) have formulated an algorithm for a least squares estimate of recharge and applied it to a hypothetical example based upon the coastal plain aquifer of Oman. Wikramaratna & Reeve (1984) extend the technique to allow more constraints on the recharge estimates, but at a very heavy price in computing costs. Neither paper treats a real aquifer, and it is probable that the errors in any real data would make such a case almost indeterminant.

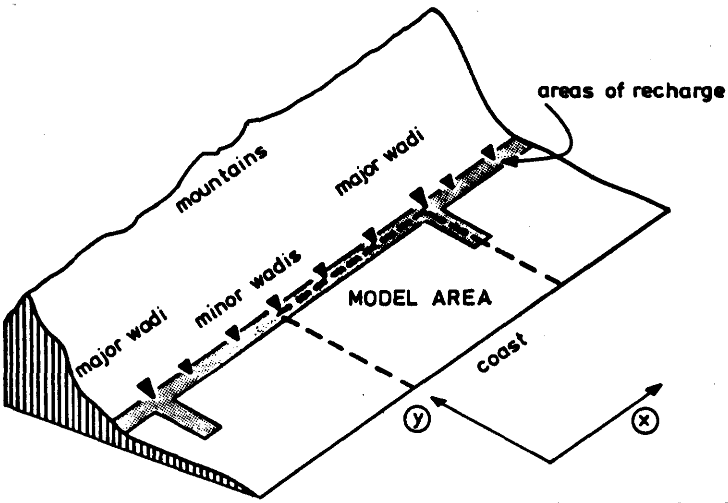


Fig. 16.3 Schematic view of an arid coastal plain aquifer, showing how recharge is constrained to certain areas (Smith & Wikramaratna, 1981)

Such inverse techniques are appealing to the computer literate, and appear to show great sophistication to the illiterate. However, their mathematical complexity conceals simple conceptual models of aquifer and recharge behaviour. They assume steady state, ignore the soil and unsaturated zones, and assume field measurements of transmissivity are applicable to regional groundwater flow. With caution, they may provide some limits to other estimates of recharge, but are not generally recommended.

16.5 Aquifer wide tracers

There are a number of ways that the concentrations of environmental tracers throughout an aquifer can be used to estimate net recharge. Although the methods are interrelated, it is useful to subdivide them into groups as follows:

(i) Mathematical transit time models

(ii) Mixing cell models of single tracers

(iii) Mixing cell models of multiple tracers

The methods differ in the mathematical formality of their approaches, data requirements and complexity of computation, as described below.

16.5.1 Mathematical transit time models

Comprehensive reviews of this approach are given by Maloszewski & Zuber (1982), Zuber (1986) and Yurtsever (Chapter 25 in IAEA, 1983). The models relate the mean age of water leaving a groundwater system to the mean transit time of a tracer through alternative mathematical models of the way the tracer moves through the system. In steady state, the mean age of water leaving the system is given by:

$$a = V/R \tag{16.9}$$

where a : mean age of water (T)
V : volume of mobile water in system (L^3)
R : recharge rate (L^3/T)

The mean transit time of a tracer is:

$$t_t = \frac{\int_0^\infty t\, C_I(t)\, dt}{\int_0^\infty C_I(t)\, dt} \tag{16.10}$$

where t_t : mean transit time of tracer (T)
t : time (T)
$C_I(t)$: concentration of the tracer at observation point due to an impulse injection (M/L^3)

Kreft & Zuber (1978) argue that $t_t = a$ only for an ideal conservative tracer which is injected and measured in flux. Concentrations in flux are obtained by sampling a spring or borehole, and can be contrasted with resident concentration obtained by analysing pore waters of rock cores.

In order to use data resulting from a continuous, and possibly varying, input concentration, we need to derive the weighting function or impulse response of the system and use the convolution integral. The impulse response could theoretically be obtained by normalising $C_I(t)$, but it is not normally possible to apply an impulse of tracer and wait long enough to observe the complete output response. Therefore a number of conceptual models have been developed, and their impulse responses theoretically determined. Details can be obtained in the references cited above.

The method consists of the following:

(i) measuring input and output concentrations of tracer over a sufficiently long period, preferably at least as long as the expected turnover time,

(ii) choosing an appropriate groundwater flow and tracer sampling model, and so the form of impulse response,

(iii) solving the convolution equation for the numerical values of the parameters of the impulse response, including a, usually by trial and error comparison of measured and predicted output values,

(iv) estimating the volume of mobile groundwater and hence the recharge rate (eqn 16.9).

The method has a number of limitations on its usefulness:

(i) a single distributed recharge is usually assumed. Maloszewski & Zuber (1982) briefly discuss two sources with different concentrations of tracer, but require additional field data to distinguish them.

(ii) If the input concentration is constant, the output will also be constant unless radioactive tracers are used to provide time dependant data.

(iii) For long transit times, the input concentrations are not well known and only a short portion of the output will have been measured. Thus determination of weighting functions and parameters will be rather imprecise.

These limitations make the method best suited to a small or rapid transit groundwater system with a single input and single output. Maloszewski & Zuber (1982) interpret several rainfall-fed springs. Allemmoz & Olive (1980) interpret tritium data in a situation where there is only recharge from ephemeral rivers.

16.5.2 Single tracer mixing cell models

The limitations of the mathematical transit time models can sometimes be overcome by using distributed, empirical models of tracer movement. The aquifer is divided into cells, based upon (i) hydrogeological units, (ii) location of recharges and discharges, (iii) presumed flow lines. The movement of tracer and water between inputs, cells and outputs is modelled using mass conservation and assuming each cell is fully mixed. The various fluxes in the model are adjusted by trial and error until the model predicts the same distribution of tracer as observed in the aquifer. Examples include Campana & Simpson (1984) and Campana & Mahin (1985).

It is common practice (although not necessary) to assume that the aquifer is in hydraulic steady state, that is there is no change in the water stored in each cell with time, in which case the mass balance for each cell is:

$$\Sigma_j \; Q_{l,j} \; + \Sigma_k \; Q_{l,k} = 0 \qquad 16.11$$

A mass balance of tracer is carried out for each cell for each time step:

$$V_l \; C_{l,t+\delta t} = \Sigma_j \; Q_{l,j} \; C_{l,j,t} \; \delta t + \Sigma_k \; Q_{l,k} \; C_{l,k,t} \; \delta t + V_l \; C_{l,t} \qquad 16.12$$

where

$Q_{l,j}$: flow into cell l from cell j; $Q_{l,j} = -Q_{j,l}$ (L^3/T)

$Q_{l,k}$: flow k into cell l from outside the aquifer (recharge, well, etc) (L^3/T)

$C_{l,j,t}$: concentration associated with flux $Q_{l,j}$ at time t (M/L^3)

$C_{l,k,t}$: concentration associated with flux $Q_{l,k}$ (M/L^3)

$C_{l,t}$: concentration in cell l at time t (M/L^3)
V_l : volume of water in cell, l (L^3)
δt : time step length (T)

The model is not based on the groundwater flow or tracer transport equations. One consequence is that the model suffers from numerical dispersion, that is the speed at which tracer spreads through the model depends partially on cell size and length of time step. The amount of numerical dispersion can be adjusted by using differing rules for computing the outflow concentration from each cell. Under the simple mixing cell rule, the inflows are mixed with the cell contents before the outflow is taken at the new concentration. The modified mixing cell rule takes the outflows before the inflows. It is possible to take intermediate positions between the two mixing rules to achieve a desired degree of dispersion.

As Maloszewski & Zuber (1982) point out, the model has a large number of parameters ($Q_{l,j}$ and V_l) which can be adjusted to achieve a match with field data. These parameters must be constrained as far as possible by other data if any reliance is to be placed on the final calibration.

All of the aquifer wide tracer models require that the tracer inputs are known for a substantial time. The mixing cell model requires them to be known for the time it takes to flush the whole aquifer, unless a set of initial concentrations can be specified for all points. For a large aquifer, this effectively restricts the model to cases where constant input concentrations and fluxes can be assumed, that is groundwater systems in a natural or under-exploited state. Good data are required on input and output concentrations and on the distribution of tracer throughout the aquifer.

16.5.3 Multiple tracer mixing cell models

The single tracer model described above can in principle be extended to many tracers simultaneously. This will make the solution overdetermined, that is there will be more than enough information to solve for intercell flows directly. However all field data contain some error, and this extra information can be used to find a "best" solution in the presence of errors.

Consider that the n cell volumes are known and there are, say, five unknown flows for each cell. If there are, say, 10 tracers then approximately 11n equations can be written to solve for the 5n unknowns.

This approach is set out by Adar & Neuman (1988) and Adar et al. (1988) who use a quadratic programming algorithm to find a weighted least squares solution for recharge to a semi-arid alluvial aquifer. For each cell, mass balances of water and tracers can be written. In matrix notation for one cell, this is:

$$\mathbf{E_n} = \mathbf{Q_n} - \mathbf{D_n} \qquad 16.13$$

A weighted mean squares criterion can be formed over all cells from eqn 16.13 as:

$$J = \sum_{m=1}^{N} E_m^T \, W \, E_m \qquad 16.14$$

where Q_m : vector of inflows
D_m : vector of outflows; n per cell, N cells
E_m : (k + 1) vector of error terms
W : (k + 1)(k + 1) diagonal matrix of weightings
J : objective function to be minimised

where $\underline{W}$ is a weighting matrix introduced to account for any available prior knowledge about errors, reliability of tracers, etc. Eqn 16.14 is subject to the constraints that all flows are positive and that outflow from one cell is equal to inflow to its neighbour.

The case modelled by Adar (1984) was of the alluvial aquifer underlying Aravaipa Creek in Arizona. He subdivided the aquifer into 5 cells, and used 14 tracers (conductivity, Mg, Ca, Na, K, HCO_3, Cl, F, Li, ^2H and ^{18}O) which were assigned weights between 0.1 and 1.0. A chemical speciation program (WATEQF) was used to check mineral saturation and hence provide one check on whether the tracers were conservative. He concluded that the model worked well provided that the sources of recharge had been correctly identified and located.

The applicability of the multiple tracer mixing cell model (as of the single tracer model) is in general to undisturbed groundwater systems or those with a low turnover time.

PART IV : CASE STUDIES

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17 INTRODUCTION TO THE CASE STUDIES

Concepts and theoretical backgrounds of techniques for estimating natural recharge have already been covered by the previous chapters. The seven case studies added to this manual serve to illustrate the practical difficulties encountered while applying the different methods. With the exception of Johansson's study they all refer to arid or semi-arid regions, and all were presented and discussed during the workshop on estimating natural recharge held in Antalya, Turkey, from 8 to 15 March 1987 (Simmers, 1988). The selection reproduced here is based principally on educational criteria and does not imply any critical appraisal of the presented results.

The cases relate to the following hydrogeological provinces described in Part II:

- alluvial basins: Rushton (northwestern India),
- sand and sandstone: Edmunds et al. (Cyprus, Sudan),
- limestone and dolostone: Bredenkamp (Transvaal, R.S.A.),
- plateau basalts: Athavale and Rangarajan (southern India),
- granitic terrain and gneisses: Athavale and Rangarajan (southern India), Thiery (Burkina Faso), Houston (Zimbabwe).

The study by Johansson refers to glacial deposits in Sweden.

The following methods, described in Part III were used:

Water balance:

- Soil moisture budgetting (Section 11.4): Rushton, Johansson, while Thiery and Houston combine this method with a lumped-parameter recharge-runoff simulation model (Section 12.6),
- baseflow analysis (Section 16.3): Houston,
- water table rise (Section 16.2): Bredenkamp, Rushton.

Darcian approaches:

- unsaturated flow (Section 11.5): Johansson (the model also describes snowmelt and soil freeze-thaw processes),
- regional numerical model (Section 12.4): Rushton.

Tracer techniques:

- signature methods (Section 11.6.2): Rushton with environmental isotopes, Athavale and Rangarajan using applied tritium,
- throughput methods (Section 11.6.3): Edmunds, Houston using chloride.

Empirical methods:

- precipitation recharge (Section 11.3): Bredenkamp.

In some of the case studies results from different methods could be compared. Houston used tracer, baseflow and soil moisture budgetting techniques, while Rushton, in his study on irrigated fields, applied a variety of methods because of the complexity of the problem.

Some authors could "calibrate" their methods. Bredenkamp was able to calibrate his empirical formula with recharge estimates collected by other methods, while Edmunds et al had tritium and lysimeter data at their disposal for comparison with the chloride concentrations at the Cyprus sites. Johansson compared the results of soil moisture budgetting and unsaturated flow analysis, both calibrated with groundwater level measurements. The case study presented by Thiery illustrates the problem of inadequate data during model calibration.

As a conclusion, it is obvious that the most reliable results are obtained if all the available data are used and a variety of methods can be applied.

18 NATURAL RECHARGE MEASUREMENTS IN THE HARD ROCK REGIONS OF SEMI-ARID INDIA USING TRITIUM INJECTION - A REVIEW

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18.1 Introduction

The semi-arid tropics (SAT) are areas where monthly rainfall exceeds potential evapotranspiration for 2 to 7 months in a year and the mean monthly temperature is above 18°C. The areas with 2 to 4.5 wet months are called dry semi-arid tropics and those with 4.5 to 7 wet months are called wet-dry semi-arid tropics.

Out of a total area of 2 x 10^6 km^2 of SAT in India, about 0.7 x 10^6 km^2 is covered by vertic soils derived from Deccan Trap Basalts and almost all the rest is covered by alfisols or related soils. Practically all of the SAT region of India is underlain by granites, basalts and indurated pre-Cambrian sediments. The undependable monsoonal rainfall is characterized by storm events and gaps and the farmer has to depend on supplemental irrigation from dug wells or shallow borewells for the sustainance of his crops.

The phreatic zone is the principal aquifer in the SAT region of India. It is located in the weathered mantle or overburden and the fracture zones in hard rocks underneath, extending to a depth of about 40m.

In view of the the importance of groundwater in supplemental irrigation in SAT agriculture, it is necessary that the exact quantum of annual replenishment (recharge) to the groundwater reserves in river catchments is evaluated, so that the safe yield is determined and the resource is properly managed and equitably distributed amonst the users.

18.2 Natural recharge measurement using injected tritium techniques

The primary source of recharge is deep percolation of a fraction of the rainfall, after the soil profile is saturated. The secondary sources of recharge are:

(i) Percolation along streams, canals and lake beds

(ii) Return flow of irrigation water, derived from surface or subsurface sources, and

(iii) Artificial recharge through spreading at suitable sites or injection in wells.

Tritium (^3H), a radioactive isotope of hydrogen, is a soft beta emitter, having mass 3 and a half-life of 12.6 years. It exists in the form of the water molecule and as such it is an ideal tracer for studying groundwater. Its use in recharge measurement involves injection of tritiated water at a certain depth in the unsaturated zone of the soil column, and study of the vertical movement that this tracer undergoes during the next hydrological cycle. The application of artificial Tritium tracer in recharge measurement is based on an assumption of the piston-flow model for water movement in the unsaturated zone. The piston flow model, proposed by Zimmerman et al. (1967a and 1967b), and by Muninch (1968a and 1968b), assumes that the percolating soil moisture moves downward in discrete layers, and any addition of a fresh layer of moisture at the surface, would push down an equal amount of water immediately below and so on, till the last such layer in the unsaturated zone will be added to the saturated regime or the water table.

18.3 Field and laboratory procedure

In the Tritium injection technique, the moisture at a certain depth in the soil profile is tagged with tritiated water. The tracer moves downward along with the infiltrating moisture due to subsequent precipitation or irrigation. A soil core is collected from the injection site after a certain interval of time and the moisture content and tracer concentration are measured from various depth intervals. The displaced position of the tracer is indicated by the peak in its concentration. The peak may be broadened because of other factors such as diffusion, irregularities in water input and streamline dispersion. The centre of gravity of the profile is assumed to correspond to the displaced position of the tagged layer. Moisture content of the soil column, between injection depth and displaced depth is the measure of recharge to groundwater over the time interval between injection of tritium and collection of soil profile.

The various catchments in which recharging measurements have been carried out are shown in Fig. 18.1. Tritium injections at representative sites in each catchment were made before the onset of monsoon, i.e., in general in the first week of June. Selection of injection sites was made on the basis of geology, soil type, topography, drainage pattern and also approachability. The number of injection sites was decided on the basis of the area and also on the feasibility of completing field and laboratory work within the specified time.

At any injection site, a relatively flat patch of land was selected, sufficiently distant from big trees, but near important landmarks such as milestones and electrical poles. The site sometimes had a thin cover of grass or shrubs. In cases where a non-agricultural site was not available, a ploughed or unploughed farm plot, having no facility for well or canal irrigation, was selected.

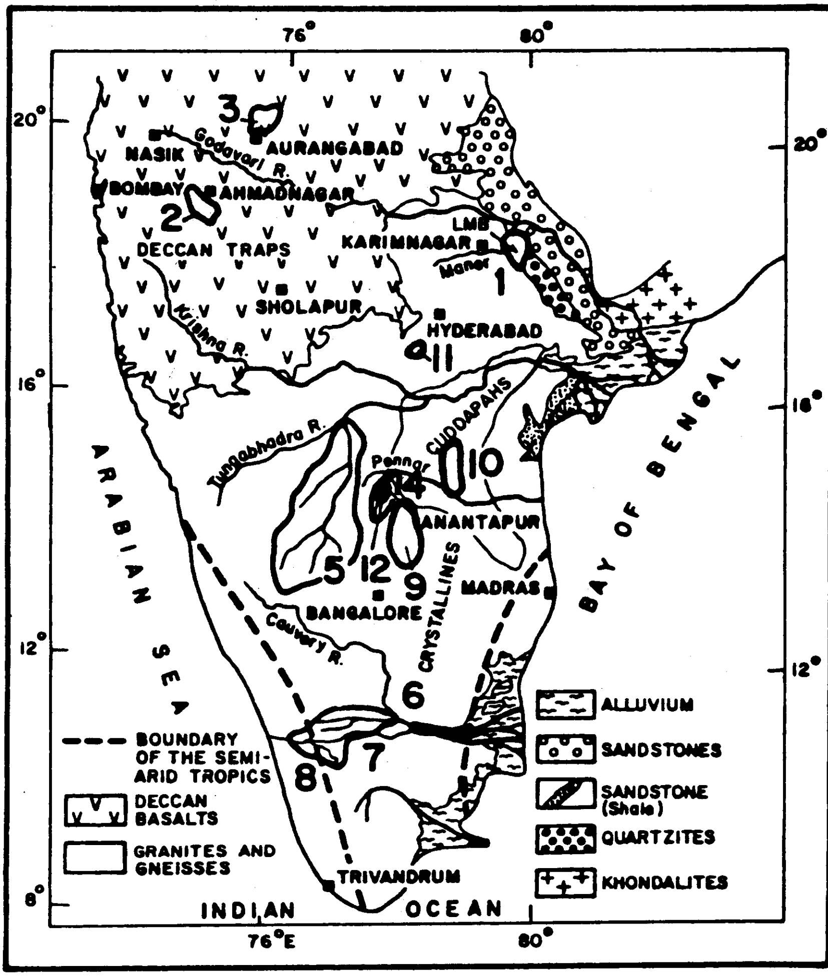

Fig. 18.1 Geological map of semi-arid south India showing recharge measurement catchments

The depth of injection varied from catchment to catchment but within the range of 60 cm to 80 cm. 2.5 ml of tritiated water, having an activity of 10 microcuries per ml was used in each injection. Five injections (total 125 microcuries of H) were made within each site. A 1.25 cm diameter hole was made down to desired depth with a drive rod and the hole backfilled with local soil after tritium injection. Four injections were placed symetrically on the circumference of a circle having a diameter of 10 cm, and the fifth injection was made at the centre of the circle. Each injection site location was precisely determined through triangulation so that it could be found easily again for vertical collection of soil profiles at the end of the monsoon. Soil profiles having a depth interval of 10 cm or 20 cm were generally collected in November-December down to a maximum depth of 3 m. Duplicate injections were made at a few sites for studying reproducibility. A second collection from a few injection points was made at the end of the hydrological cycle, to study the effect of evaporation and transpiration during the post-monsoon lean period.

Table 18.1 Mean values of precipitation recharge for hard rock covered catchments in semi-arid India

Sl. No.	Basin/Watershed name	Coordinates (Lat & Longt)	Main rock types	Area (Sq. Km.)	Year of measurement	No. of recharge measurements	Average rainfall (mm)	Mean recharge (mm)
1.	Lower Maner	18° 05' -18° 42'N 79° 32' -80° E	Sand Stone, shale, quartzite, granite	1,600	1976	26	1,250	100
2.	Godavari-Purna	19° 45' -20° 15' N 75° 10' -75° 50' E	Basalt	1,091	1980	24	652	56
3.	Kukadi	18° 45' -19° 10' N 74° 10' -74° 40' E	Basalt	1,153	1980	19	612	46
4.	Marvanka	14° 15' -15° N 77° 15' -77° 47' E	Granite, greiss, schist	2,044	1979	19	550	42
5.	Vedavati							
	(1) West Suvarnamukhi	13° 15' -13° 45' N 76° 30' -76° 45' E	Granite, gneiss schist	958	1978	18	565	39
	(2) Lower Hagari	14° 45' -16° N 76° 45' -77° 30 E	Granite, gneiss, schist	3,679	1978	44	565	6.5
6.	Noyil	10° 54' -11° 19' N 76° 39' -77° 56' E	Granite, gneiss, schist	3,420	1979	21	715	69
7.	Vattamalaikarai	10° 52' -11° 52' N 77° 15' -77° 45' E	Granite, gneiss, schist	512	1979	2	460	61
8.	Ponnani	10° 15' -11° N 76° 15' -77° 15' E	Granite, gneiss, schist	3,973	1979	9	1,320	61
9.	Chitravati	13° 35' -14° 50' N 77° 30' -78° 15' E	Granite, gneiss, schist	6,100	1981	48	615	25
10.	Kunderu	14° 30' -16° N 77° 45' -79° E	SST, Shale, LST, quartziti, granite, gneiss	8,650	1982	45	615	29
11.	Anrepalle Watershed (Mahaboobnagar)	17° 47' -17° 53' N 78° 33' -78° 43' E	Granite	64	1984 1985	15 12	563 583	32 17
12.	Manila Watershed (Anantapur)	14° 30' -14° 36' N 77° 38' -77° 44' E	Granite & gneiss	40	1986	25	390	24

Note : Serial numbers correspond with numbers in Fig. 1

In the laboratory, the moisture content in about 25 g of soil, taken from each sample, was determined on a torsion balance heated by an infrared lamp up to 105°C. The rest of the sample was subjected to partial vacuum distillation. 4 ml of the distillate was mixed with 10 ml of scintillator solution and the Tritium activity was counted using a liquid scintillation spectrophotometer. These field and laboratory procedures are described in detail by Athavale et al. (1978 and 1980).

The tritium activity of each 10 or 20 cm depth interval of a profile was plotted against depth along with moisture content. Grain size analysis of the soil core material was carried out.

18.4 Results and discussion

The results of recharge measurements carried out in various catchments in the SAT region of South India are presented in Table 18.1. Some of the results are already published by Athavale et al. 1980 (for Sl. no. 1 in Table 18.1), 1983a (for Sl. no. 2 and 3) and 1983b (for Sl. no. 4). The work in Vedaveti catchment (Sl. no. 5) and in Noyil, Ponnani and Vattamalaikarai catchments (Sl. no. 6,7,8,) was carried out at the request of the Central Ground Water Board of Govt. of India and the results have been communicated to them. Rangarajan and Ramesh Chand (1987) have carried out recharge experiments for Chitravati and Kunderu catchments.

As can be seen from this table, the area of these catchments varies from 500 km² to 8500 km². On average, each recharge measurement site represents an area of 50 to 200 km².

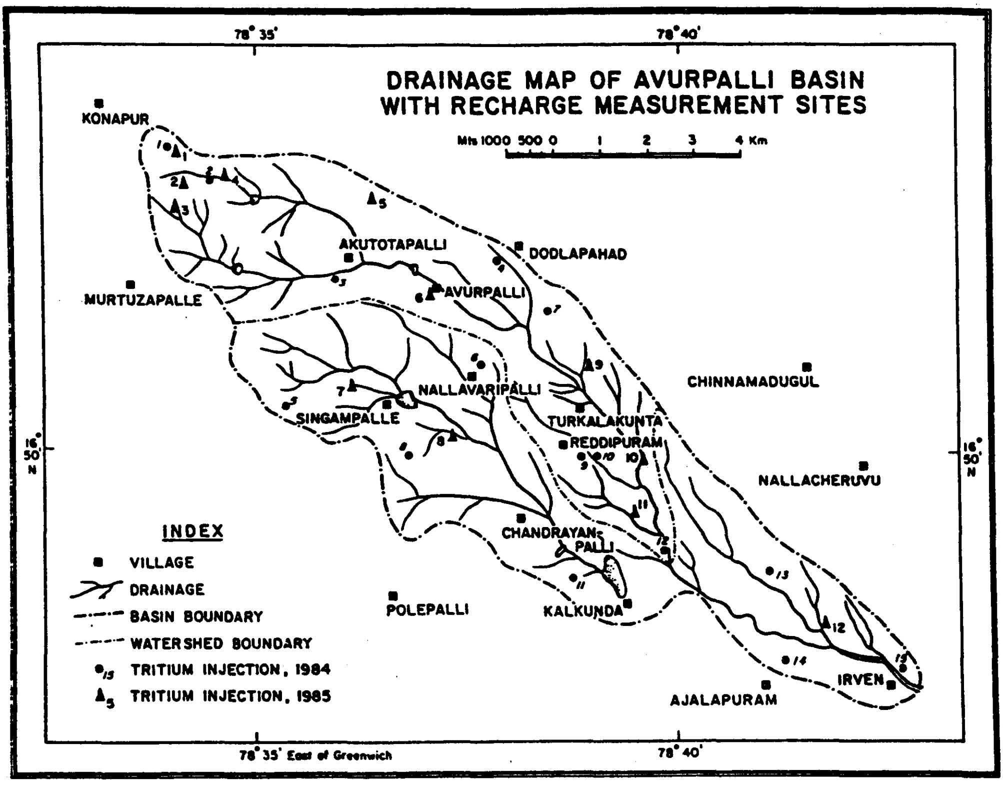

Fig. 18.2 Aurepalle catchment

Recently some detailed studies on two small catchments (Sl. no. 11 and 12) have been carried out. In the case of these catchments the density of recharge measurement was one site per 2 to 4 km². In these cases the measurements have been carried out for two monsoons (Rangajaran et al., 1987; Rangajaran & Ramesh Chand, 1987b). Results of recharge measurements for Aurepalle catchment for 1984 and 1985 are reported here in detail as an illustration. In the case of Manilla catchment, 1985 results are presented and the samples collected in November 1986 are being processed.

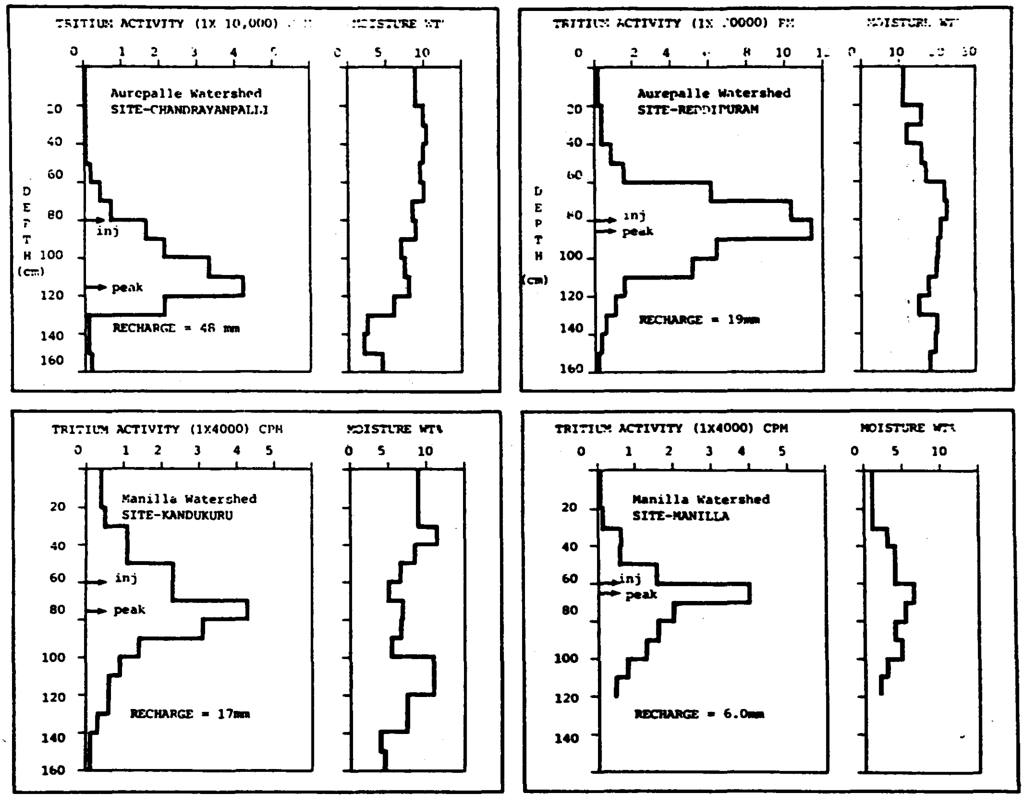

Fig. 18.3 Tritium and moisture content variation with depth at 4 injection sites in Aurepalle catchment

18.5 Recharge measurements in Aurepalle catchment

Tritium injection studies were carried out in the Aurepalle catchment during two hydrological cycles, 1984 and 1985. This catchment, covering an area of approximately 64 km² is located in granites and has a cover of alfisols. The weathered zone thickness varies from a few metres up to 20 metres, with an average value of about 8-10 m. Recharge measurements at 15 sites injected in June 1984 gave a mean value of 31.9 mm, which is 6.2% of the 1984 annual rainfall of 563 mm. Input to groundwater system due to direct percolation of precipitation, was estimated as about 2 x 10^6 m^3.

Recharge measurements were repeated in the 1985 hydrological cycle. The mean value of recharge measured at 12 sites was found to be 17 mm, which is only 3% of the total rainfall of 583 mm.

Fig. 18.2 shows a drainage map of Aurepalle catchment and tritium injection sites for 1984 and 1985. Typical tritium vs. depth profiles in the case of four injection sites from this catchment are presented in Fig. 18.3. The recharge data for 1984 and 1985 are presented in Table 18.2.

Table 18.2 Recharge due to monsoon precipitation in Aurepalle catchment during 1984 and 1985

1984			1985		
Sl no.	Site	Recharge (mm)	Sl no.	Site	Recharge (mm)
1	near Konapur	-2	1	near Konapur	5
2	Akuthotapally-Konapur	-15	2	Lambaditanda	22
3	S of Akuthotapally	39	3	Akuthotapally-Konapur	8
4	Dodlapahad	10	4	Akuthotapally-Konapur	19
5	Singampally II	43	5	N of Akuthotapally	4
6	Nallavarpally	70	6	Aurepalle	17
7	Aurepalle-Chinnamadgul	103	7	Singampally	33
8	Singampally I	78	8	S of Singampally	39
9	Reddipuram	16	9	Turkalkunta	25
10	Reddipuram (repeat)	16	10	Reddipuram-Chinnamadgul	8
11	Chandrayanpalli-Kalkunda	37	11	Reddipuram-Kalkunda	15
12	Kalkunda NE	8	12	near Irven	12
13	Ajalapuram-Chinnamadgul	5			
14	near Ajalapuram	31			
15	Irven	24			
Mean of 14 sites		**31.9**	**Mean of 12 sites**		**17.3**

18.6 Effect of rainfall pattern on recharge

Infiltration and runoff due to a precipitation event depend upon the amount and intensity of rainfall as well as the antecedent moisture condition of the soil profile. The cumulative recharge due to a monsoonal season, made of several precipitation events, is also considerably dependent on the characteristics of the events as well as their spacing. This is well brought out in the study of precipitation recharge over the two hydrological cycles 1984 and 1985, in the case of the Aurepalle catchment.

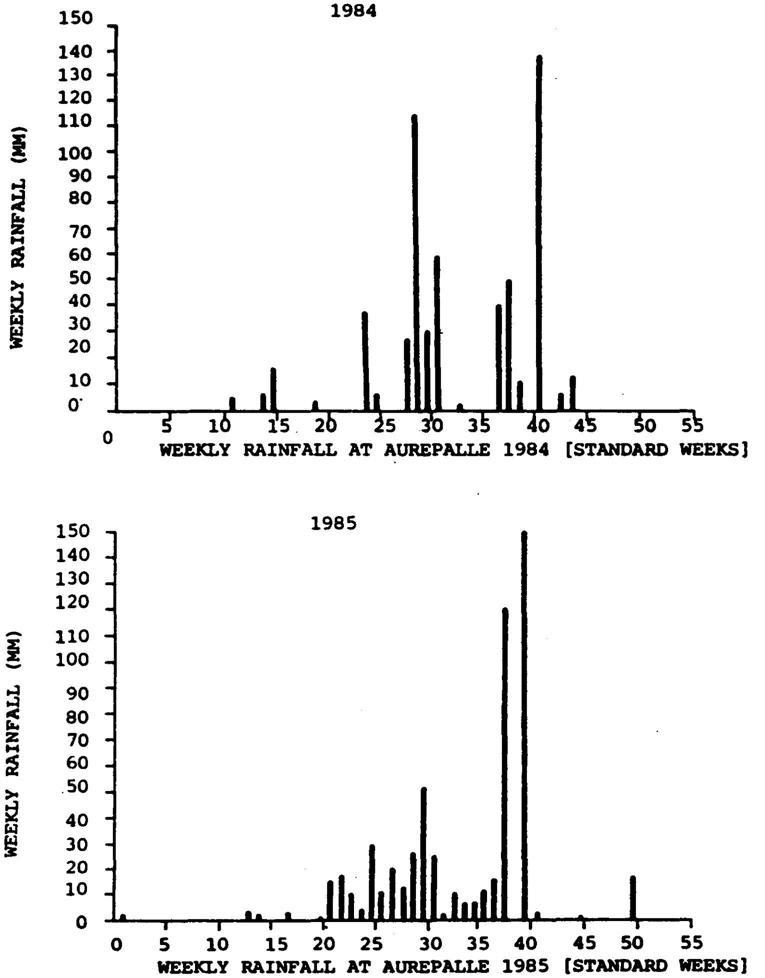

Fig. 18.4 Rainfall pattern in Aurepalle catchment

The annual rainfall values for 1984 (563 mm) and for 1985 (583 mm) are almost equal but there is a great deal of difference in the rainfall pattern of the two consecutive years, as can be seen from Fig. 18.4. The 1984 monsoon is characterised by two major storm events, occurring during the 29th and 41st standard weeks. These events resulted in a rainfall of 115.6 mm and 139.6 mm respectively and accounted for nearly 45% of the annual rainfall. The mean recharge value for 1984 was 32 mm. In the case of the 1985 monsoon, two storm events totalling 120.1 mm and 149.4 mm also accounted for 46% of the total rainfall. However, these events occurred during the 38th and 40th standard week, a situation giving rise to higher runoff than the 1984 storm events, which were separated by about 11 weeks. Thus the average recharge due to the 1985 monsoon (17 mm) is lower than due to 1984 (32 mm) by about 50%, as a result of the rainfall pattern. Similarly, the recharge would be more than the 32 mm average figure, if the same amount of precipitation is more evenly distributed than in 1984.

This case illustrates the need for generating recharge data in a small catchment for a period of about 5 years and for preparing a rainfall pattern:recharge model, which would then be able to predict the recharge directly from rainfall data.

18.7 Effect of grain size on recharge values

A mechanical grain size analysis of the soil material collected beneath the injection depth was carried out in the case of all catchments. The soil was soaked in water in a beaker for about a day. The wet sample was first washed with a standard sieve of size 60 µm so that all particles less than 60 µm in size and representing the silt-clay fractions were removed. The remaining sample was dried in an oven and further sieved into the sand (60 µm to 2 mm mesh) and gravel (> 2 mm mesh) fractions (Athavale et al., 1978).

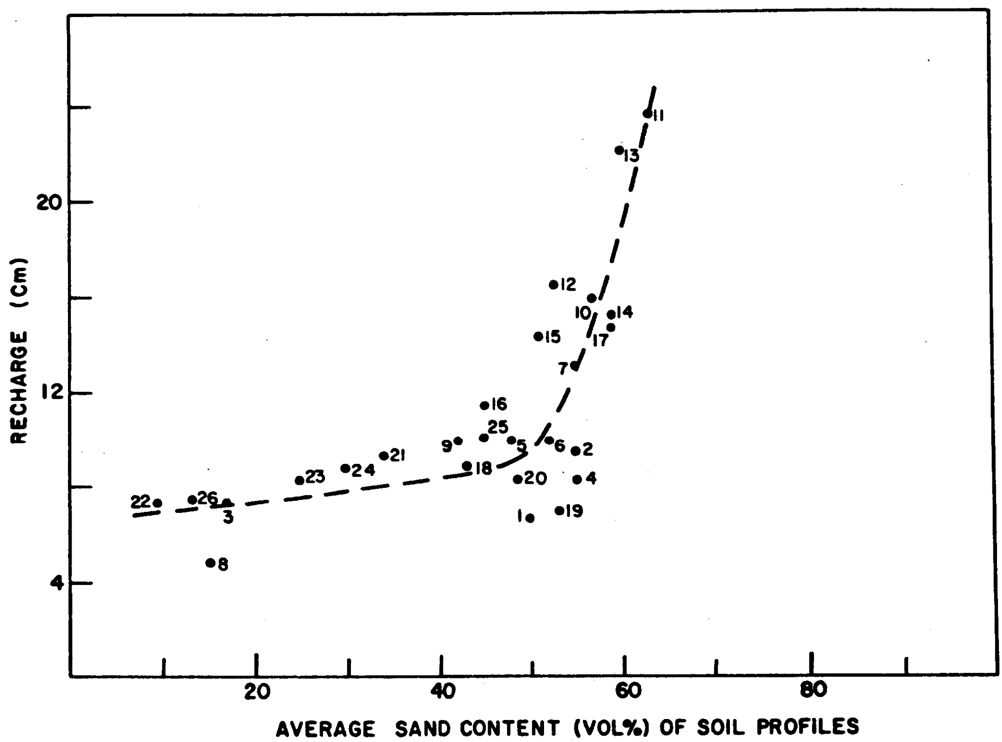

Fig. 18.5 Variation of recharge with average sand content of soil profiles

The recharge values of individual sites were then plotted against the average sand content (volume percent) below the depth of injection. An example of such a study for 26 sites in the Lower Maner catchment (Sl. no. 1 of Table 18.1) is shown in Figure 18.5. A best fit curve drawn through the points has one asymptote at about 70%. A more or less linear correlation between recharge values and sand content seems to be valid for two separate segments of the curve. One of these is in the sand percentage range of 20-40 and the other is in the range 55-65%.

18.8 Diffusion of injected Tritium activity in soil and variability of recharge rates

In order to study the extent of lateral spreading of injected tritium activity in the soil due to diffusion, an experiment was conducted in clayey soils at one site in Lower Maner Catchment (no. 1 in Fig. 18.1 and Table 18.1). Soil cores for this study were obtained from the point where Tritium was originally injected and also from points at distances of 0.5, 1.0 and 2.0 m from the injection point. The tracer was injected in April, 1976 and samples were collected in July, 1977 (Athavale et al., 1978). Tritium activity was present up to a lateral distance of 1.0 m, while it was beyond the detection limit at the distance of 2.0 m. The cumulative Tritium activity of profiles at distances of 50 cm or 100 cm was found to be 0.26 and 0.005 times the activity at the injection point.

The values fit into the exponential law of diffusion. For the experiment lasting for 15 months (t = 450 days) the diffusion constant for clayey soil at site Dhawanda was found to be 1.21×10^{-5} $cm^2 s^{-1}$. This is within the range of reported values for diffusion rates in clays.

In the case of the Tritium injection method, we have measured the reproducibility error as ± 10%. Availability of such estimates of experimental error in the case of other methods used for recharge estimation is desirable.

Table 18.1 presents data on deep percolation or recharge in the case of 12 catchments, covered by vertic soils and alfisols and underlain by basalts, granites, gneisses, schists, quarzites, sandstones and shales. The annual rainfall values range from 390 mm to 1250 mm and the average recharge values from 6.5 mm to 100 mm. A broad correlation between the rainfall and recharge amounts exists although this correlation is greatly influenced by other factors such as soil type, topography and hydrogeological conditions, in addition to the annual effect of rainfall pattern, as shown in the case of the Aurepalle catchment.

The cumulative effect of the various factors leading to scatter of recharge values can also, at best, be expressed as standard errors of the average value. In the case of Aurepalle catchment (Table 18.2) the values are 31.9 ± 33.0 and 17.3 ± 11.0.

18.9 Conclusions

The Tritium injection technique is found to be a comparatively quick and reliable method in estimating groundwater recharge due to precipitation having a peaked monsoonal character in the semi-arid tropical region. These recharge values are in good agreement with those obtained using various other techniques. Recharge values in a catchment are greatly influenced by the pattern of seasonal rainfall. There is, therefore, a need for conducting recharge measurements for a period of about 5 years in representative catchments covered by alfisols and vertisols, in order to prepare a data based model for direct determination of recharge from the rainfall record for any specified year. Systematic recharge measurements over large areas would provide a useful basis for optimal utilisation of the replenishable but limited groundwater reserves of semi-arid tropical regions. The recharge measurements will also form a primary data base for artificial recharge programs which would be needed for meeting the increasing demand for groundwater by farmers of the semi-arid tropical regions.

19 QUANTITATIVE ESTIMATION OF GROUNDWATER RECHARGE BY MEANS OF A SIMPLE RAINFALL-RECHARGE RELATIONSHIP

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19.1 Introduction

Since 1963 the Bo Mopolo dolomite region (Fig. 19.1) has been studied with the object of estimating groundwater recharge and to determine the exploitation potential of different dolomite compartments.

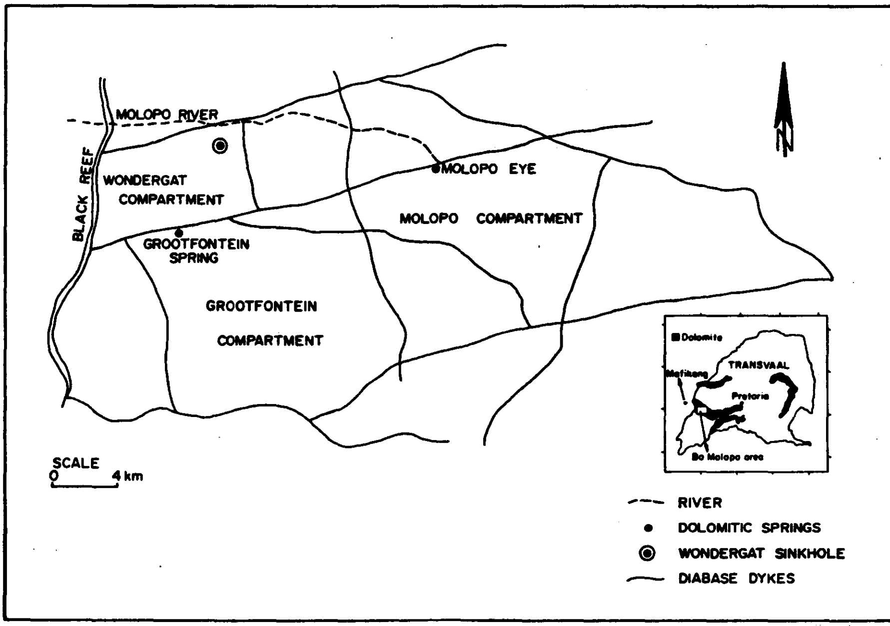

Fig. 19.1 Map showing some of the Bo Molopo dolomite compartments, South Africa

The Grootfontein compartment, recently supplemented by the Molopo eye is the main water supply to Mafekeng/Mmabatho. Several methods were employed to estimate its groundwater recharge, e.g.

- water balances
- use of natural isotopes and tritium profiles
- aquifer modelling
- rainfall-recharge equations

This paper deals with the application of a simple recharge equation which provides estimates of the variability and average groundwater recharge. This is demonstrated by comparing results with estimates that were derived by other methods as well as in the reconstruction of the discharge of dolomitic springs from annual recharge estimates.

19.2 Rainfall recharge equation

19.2.1 Development

The development of a recharge equation was initially attempted using daily rainfall and climatic data (Bredenkamp et al., 1974) but was superseded by models based on monthly input data (Bredenkamp, 1978) with a substantial improvement in the simulation. This prompted annual models to be tested yielding equally good and often better results.

The outcome of an evaluation of different rainfall recharge equations provided the following simple relationship:

$$RE(I) = A(RF(I) - B) - SMD \qquad 19.1$$

where RE(I) denotes the recharge for year I; RF(I) is the rainfall for year I; A,B are lumped parameters to be optimized and SMD represents the accumulated soil moisture deficit which is also an integrated parameter.

B represents the threshold rainfall which has to be exceeded to effect recharge. Parameter A determines the proportion of the excess that constitutes recharge whilst B also determines the variability. Any negative outcome of eqn 19.1, i.e. rainfall below B, is regarded as soil moisture deficit, but this is only allowed to accumulate to a set maximum value C. The optimisation however showed that for most instances C = 0 implying that soil moisture deficit can be disregarded in the dolomite if annual rainfall values are used to simulate recharge.

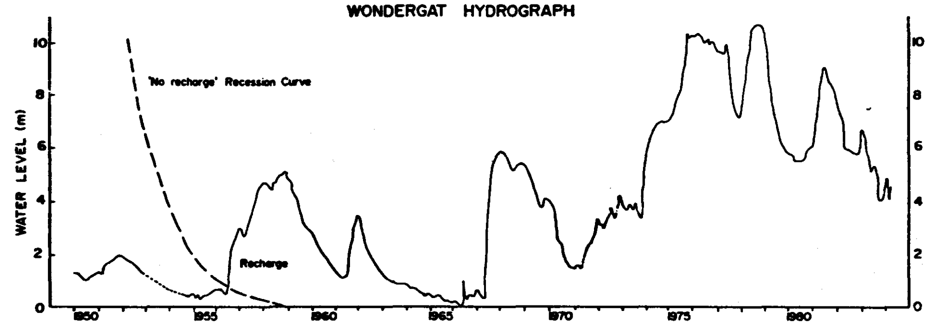

Fig. 19.2 Hydrograph of the Wondergat sinkhole which was used to reconstruct annual recharge

19.2.2 Reference recharge values

The groundwater hydrograph of the Wondergat sinkhole (Fig. 19.2) was used to reconstruct annual values of recharge as an effective rise of the water table, according to the method proposed by Bredenkamp et al. (1974) and Bredenkamp (1978).

By incorporating the "no recharge" recession, the difference between the actual water level and the level to which it would have declined, had no recharge taken place, was assumed to present the recharge. The "no recharge" response of the Wondergat was composed of the 1960/1 and 1979/80 recession limbs of the hydrographs. Although monthly recharge equivalents were obtained they were combined to yield annual recharge values. This was done to overcome the rainfall recharge lag and because annual estimates of recharge are adequate for most quantitative groundwater assessments.

The Wondergat annual recharge equivalents are shown in Fig. 19.3. These assess their ability to achieve good simulation. The difference between the reference and simulated values were squared and summed (F value in Table 19.1) to compare the performance of models. The correlation coefficient reflects the degree of correspondence attained.

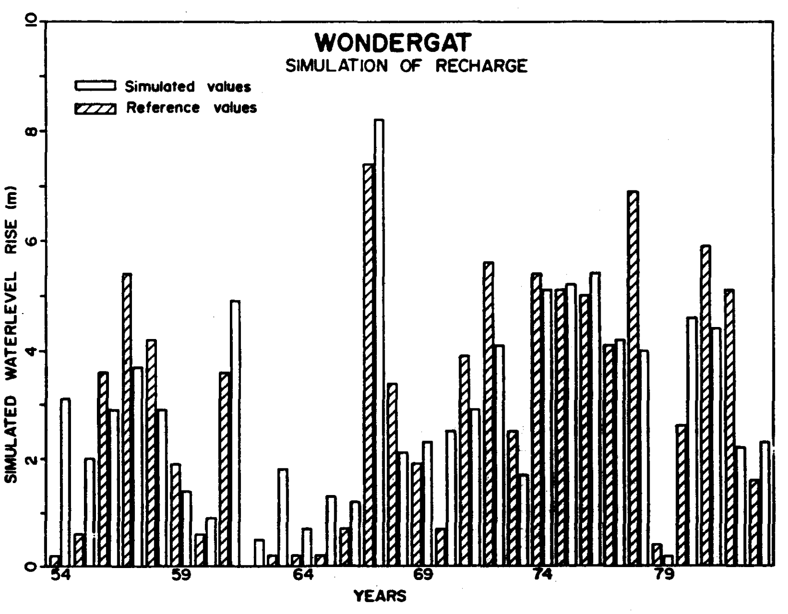

Fig. 19.3 Recharge values derived from annual rainfall and reference recharge values reconstructed as a water level rise

19.2.3 Optimisation

Although convergence of the F-value was sensitive to variation of B in eqn 19.1, it was hardly affected by adjustment of A. A unique solution of A could only be achieved if either porosity of the aquifer is known or if an independent estimate of the average recharge is available.

A finite element model study of the Grootfontein aquifer indicated a porosity of about 0.028. Using this porosity value A and B were found by optimisation to be 0.35 and 360 respectively (Table 19.1). However values for A of 0.32-0.36 and B of 340-360 also produced acceptable simulations.

Table 19.1 Convergence of F for different values of A, B and C in recharge formula, eqn 19.1

A	B	C	Simulated recharge (mm)	Residuals squared F
Annual version				
0.30	320.0	0	83.0	58.58
0.30	340.0	0	77.0	59.90
0.30	360.0	0	71.0	63.98
0.30	380.0	0	65.1	70.74
0.30	400.0	0	59.3	80.07
0.35	320.0	0	96.8	63.65
0.35	340.0	0	89.8	58.05
0.35	360.0	0	82.8	56.20
0.35	380.0	0	75.9	58.01
0.35	400.0	0	69.2	63.35
0.40	320.0	0	110.7	87.19
0.40	340.0	0	102.7	72.62
0.40	360.0	0	94.7	62.96
0.40	380.0	0	86.8	58.09
0.40	400.0	0	79.0	57.84
0.35	360.0	10	82.8	56.20
0.35	360.0	20	82.8	56.20
Monthly version				
0.40	60.0	0	81.7	65.2

Notes. Porosity = 0.028 in all cases. Correlation coefficient in range 0.794 - 0.796.

In the optimisation of the monthly version of eqn 19.1, a value for A = 0.40 and B = 60 was obtained. As is shown in Table 19.1 (F-value) the monthly model is almost as good as the annual version and the differences between simulated recharge values for monthly and annual simulations are surprisingly small as is indicated in Fig. 19.4. Fig. 19.3 shows the comparison between simulated and reference recharge values using annual rainfall in eqn 19.2.

19.3 Model verification

Substituting the values of A and B that were obtained by optimisation into eqn 19.1 gives:

$$RE(I) = 0.35\ (RF(I) - 360) \qquad 19.2$$

Validation of this rainfall recharge relationship and its general application were sought by verifying results with recharge estimates obtained by other methods.

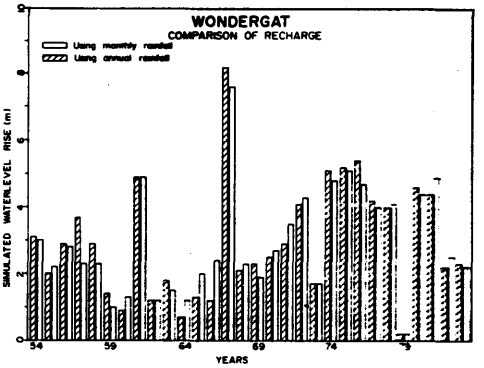

Fig. 19.4 Comparison of recharge values obtained using annual and monthly rainfall in eqn 19.2

19.3.1 Water balance calculations

The boundaries of the Grootfontein compartment (165 km²) are well delineated, and influent and effluent leakage can be assumed to be equal. This allows groundwater recharge to be estimated using a simple water balance over two periods (see Table 19.2).

Table 19.2 Recharge to Grootfontein compartment using a simplified water balance

Period	Pumpage ($10^6 m^3$)	Change in storage* ($10^6 m^3$)	Rainfall (mm)	Calculated recharge (mm)
Sep.80-Aug.82	27.0	0	735+467	164
Sep.82-Aug.83	13.5	6.26	480	48

*assuming aquifer porosity 0.028

Using these recharge and rainfall values yields A = 0.285 and B = 311.

The response of the Grootfontein aquifer has been successfully simulated by a finite element model (FEM) (Van Rensburg, 1985) and although further refinement is to be carried out the following recharge values were obtained from a water balance incorporating all grid elements (Table 19.3).

Using the recharge value in column 3 yields A = 0.227 and B = 323, which is in close agreement with the previous estimates of these parameters.

Table 19.3 Comparison of recharge estimates by FEM and eqn 19.2 (mm)

Hydrological year	Rainfall	Recharge* using FEM	Recharge using eqn 19.2
1980/81	735	114	131
1981/82	467	42.5	37
1982/83	480	36	42
1983/84	314	-2.3	-16

*assuming aquifer porosity 0.028

The different sets of values obtained for A and B are not very critical as is indicated by the following comparison of recharge estimates (Table 19.4). There is good correspondence, except for rainfall values close to the threshold value B.

Table 19.4 Comparison of recharge estimates using different constants in eqn 19.1

Rainfall (mm)	Recharge (mm) (A = 0.35; B = 360)	Recharge (mm) (A = 0.285; B = 310)
400	14	25.6
500	49	54
600	84	82.6
700	119	111

The variability in recharge that is not accounted for by eqn 19.2 is not surprising if one considers variability of rainfall intensity, its duration and spatial distribution, and the heterogeneity of the recharge area.

Attempts to obtain a better simulation by weighting A and B with climatic indices, such as the ratio of actual evaporation to average evaporation and days of rainfall, did not effect a significant improvement.

19.3.2 Reconstruction of the flow of dolomite springs

By means of the annual estimates of recharge that were derived from the recharge equation, the flow of a number of dolomitic springs was reconstructed according to the method proposed by Bredenkamp (1987).

The method employed is that proposed to recontruct the flow of the springs at Pretoria Fountains (Bredenkamp et al. 1985). The average effective area necessary to sustain the spring flow is calculated by means of the average recharge:

$$Q = RE \times Area \qquad 19.3$$

where RE is the average recharge and Q the average flow of the spring.

The calculated average area enables the annual recharge values, estimated from eqn 19.2, to be converted to annual flow rates. These were distributed incorporating carry-over of flow from the preceding years. In the case of Maloney's Eye which is a prominent dolomitic spring west of Krugersdorp, the following relationship produced a good simulation of the spring flow:

$$Q(I) = 0.333(2Q(I-1) + 0.3QR(I-1) + 0.3QR(I-2))$$

where Q(I) is the flow for year I and QR(I) is the recharge converted to an annual flow rate by means of eqation 19.3.

Fig. 19.5 shows the measured and simulated flow rates for this spring and although 1:1 correspondence is not obtained, fluctuations are synchronous for most of the record. The simulated flow during the high rainfall period of 1976-78 is significantly lower than the measured values, possibly indicating an addition of groundwater from adjacent compartments during the years of excessive rainfall.

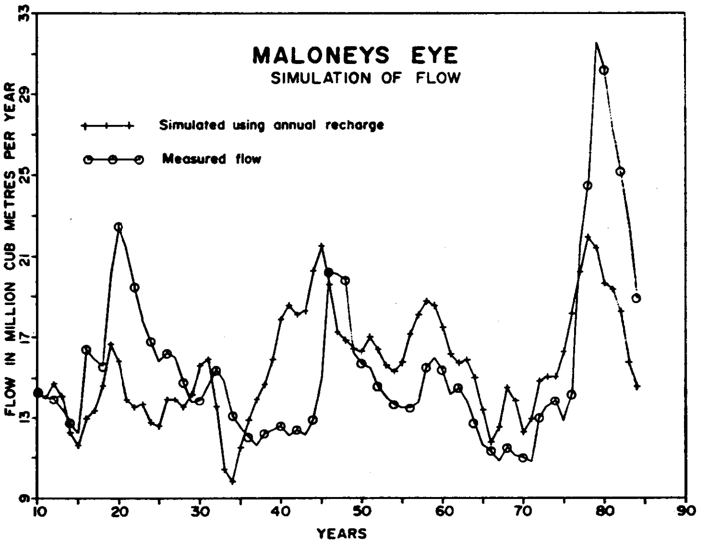

Fig. 19.5 Comparison between the reconstructed and measured flows of Maloney's Eye

In a more refined study the simulation of several dolomitic springs was carried out by Bredenkamp and Zwarts (1987).

19.4 General rainfall-recharge equation

Quantitative estimates of the average annual recharge which had been obtained for different dolomite regions were used to verify the applicability of eqn 19.2 in estimating the average annual recharge for different areas. A comparison of these estimates (Table 19.5) reflects a high degree of correspondence.

Table 19.5 Comparison of recharge for different areas in relation to estimates using eqn 19.2

Locality	Catchment area km^2	Long-term Av. Rainfall mm	Recharge (Eq. 2) mm	Recharge using other methods (mm)	Methods used for recharge estimation (Bredenkamp et-al 1987)
1. Zuurbekom	130	672	109	104	Ground-water balance and chemical balance
2. Gemsbokfontein	64	665	107	86	Water balance over two periods
3. Bank	154	640	98	109	Hill method; Average flow of spring; Inferred from spring flow
4. Steenkoppies	177	630	94,5	103	Average flow of spring; Darcy equation
5. Holfontein	75	630	94,5	93,5	Water balance
6. Kliprivier	470	700	119	134	Based on 2 mm recharge per rainfall event
7. Grootfontein	165	560	71	65	Water balance
8. Grootfontein	165	546	65	69,5	Water balance finite element model
9. Molopo eye	-	590	77	87	Reconstructed flow of spring
10. Pretoria/Rietvlei	30	682	113	104	Water balance
11. Sishen	-	390	11	31	Finite element simulation
12. Sabie	33,4	1 250	311	287	Based on flow from drainage tunnel
13. Pretoria Fountains	30	675	110	108	Average flow of Fountains-east springs
14. Kuruman	-	346	0	11	Average flow of Kuruman eye

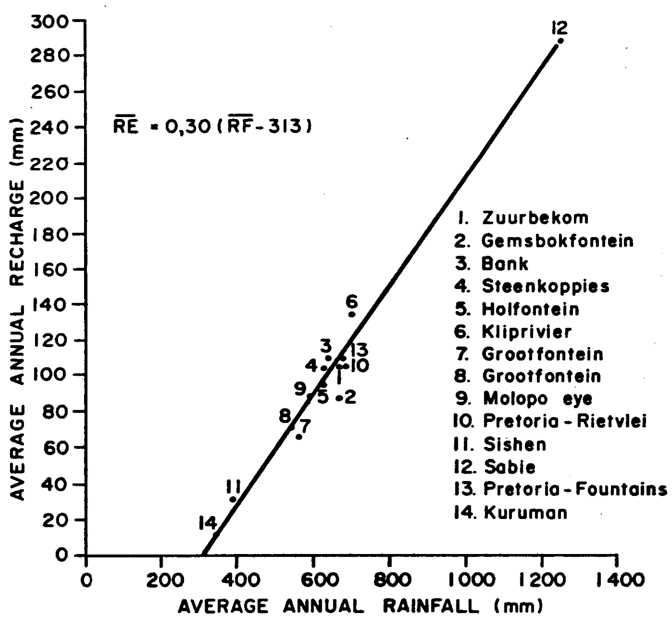

Fig. 19.6 Comparison of average annual recharge estimates for different regions with annual precipitation

However by simply plotting the recharge estimates, (column 4 in Table 19.5) against average annual precipitation

(Fig. 19.6) a linear relationship with the following values for A and B is obtained:

$$RE = \underset{A}{0.30}\ (RF - \underset{B}{313}) \qquad 19.4$$

where RE is the average annual recharge and RF denotes the average annual rainfall.

The high correlation coefficient ($r = 0.989$) reflects excellent agreement between the recontructed and reference recharge values. This indicates that eqn 19.4 rather than eqn 19.2 is the general rainfall-recharge relationship for all dolomite areas in the summer rainfall regions of South Africa.

19.5 Linear rainfall-recharge relationship for non-dolomitic aquifers

In a similar quantitative study of groundwater recharge for a shale aquifer in the Pretoria-Rietondale area (Bredenkamp, 1987), annual recharge values were derived from an interpretation of the hydrographs of three monitoring boreholes, again according to the method that was used for the Wondergat (c.f. section 19.2.2).

As is indicated by Fig. 19.7 a linear relationship between rainfall and recharge once more emerges. The recharge equation which is obtained by taking the average of three hydrograph interpretations then becomes:

$$RE(I) = 0.20\ (RF(I) - 395) \qquad 19.5$$

which indicates that 20% of rainfall in excess of 395 mm constitutes recharge for this shale aquifer as compared to 30% of rainfall in excess of 313 mm for dolomite aquifers.

Too few quantitative recharge estimates for non-dolomitic aquifers are available to check if eqn 19.5 could also be adopted on a regional basis, but from the results that were obtained for the dolomite this seems justifiable.

19.6 Conclusions and recommendations

For dolomitic areas in the summer rainfall regions of South Africa, the average recharge as well as the annual variability of recharge, could be estimated by means of eqn 19.4.

A similar linear relationship, however, with an adjusted value for slope (A) and threshold value (B), was found to apply in the case of a shale aquifer as well.

The method shows great potential as an easy means of estimating recharge which is often difficult if not impossible to obtain reliably by other techniques. Ascertaining the values of A and B should be pursued for different areas and various types of aquifers. It seems more than likely that equations similar to 19.4 and 19.5 would be forthcoming for other regions of the world.

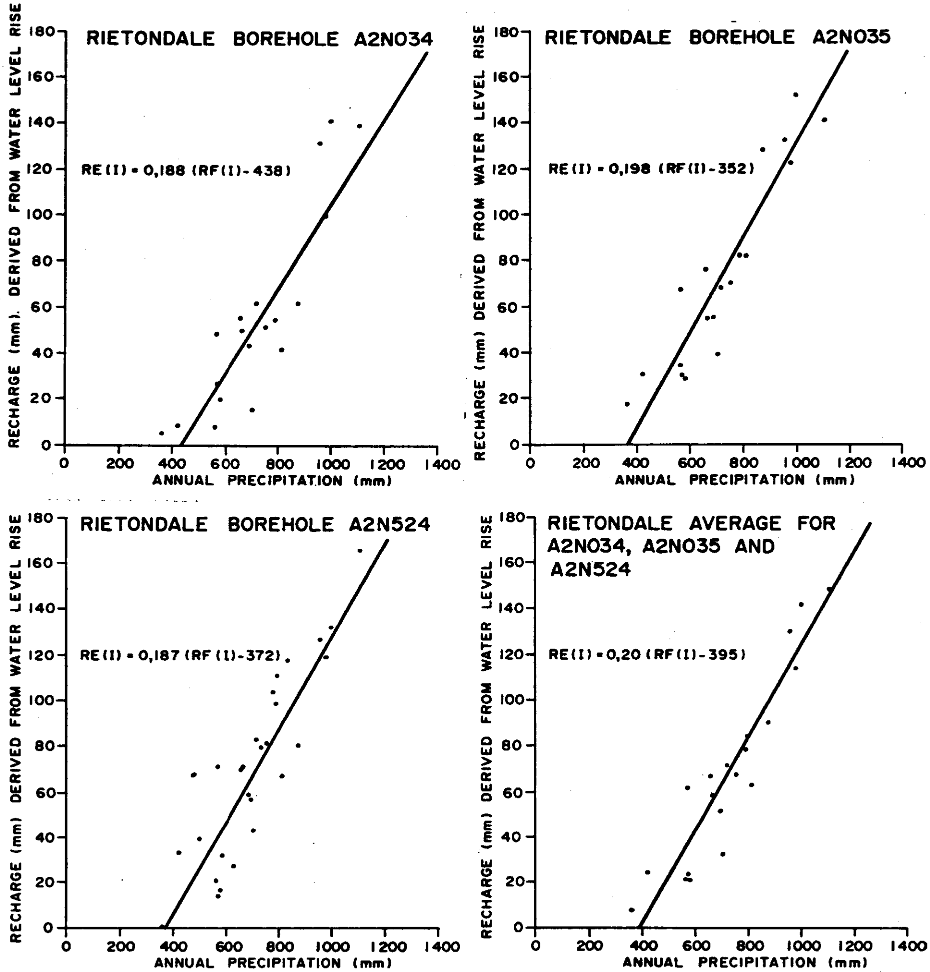

Fig. 19.7 Relationship between annual recharge and rainfall for monitoring boreholes in the Pretoria-Rietondale area

20 SOLUTE PROFILE TECHNIQUES FOR RECHARGE ESTIMATION IN SEMI-ARID AND ARID TERRAIN

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20.1 Introduction

Conventional methods for recharge estimation have limitations when applied to arid and semi-arid regions; the use of Tritium profiles is also not always applicable. Unsaturated zone solute profiles, using a reference solute such as chloride, offer an alternative technique. Sampling may be undertaken by percussion drilling, augering or from dug wells; the methods developed are described and examples discussed. Recharge estimates using chloride profiles from Cyprus (420 mm mean annual rainfall) are in good agreement with results estimated from Tritium profiles and indicate a mean annual recharge of around 50 mm/y. In Central Sudan (180 mm mean annual rainfall), good agreement was found between adjacent unsaturated zone chloride profiles and these indicated a net annual direct recharge via interfluve areas of around 1 mm/y. It is concluded that solute profiles offer a cheap and effective tool for estimating direct recharge in porous lithologies of semi-arid regions and also for investigating recharge history, providing input data for chloride are available. In more arid regions, however, a component of discharge may occur during hyperarid episodes. Further validation of moisture composition using stable isotope techniques is required under such conditions.

20.2 Model for the use of solutes to estimate recharge

A working hypothesis for the use of solutes to evaluate the direct component of recharge is illustrated in Fig. 20.1.

The input of solutes to the aquifer depends initially on the total atmospheric fallout per unit time, made up of rainfall (F_p) and the net dry deposition (F_d) fluxes. Both the rainfall amount (P) and local composition of the total deposition (F_p + F_d) may be determined for a given site, although the regional variation in both quantities must be considered if recharge estimation is required for large areas.

Solutes will be dispersed on and transported through the upper soil during the rainy season at rates depending on the rainfall intensity. These solutes will undergo concentration as a result of evapotranspiration (E). Solutes may be removed from solution by plant uptake, by mineral precipitation, or by adsorption. Similarly solutes may be released by decay of dead plant material, by mineral dissolution or by desorption. In the absence of a knowledge of these ancillary fluxes, only those constituents for which there is no <u>net</u> release or storage by the soil or rock matrix may be used. Chloride is the solute that is most frequently used and most conveniently meets these requirements.

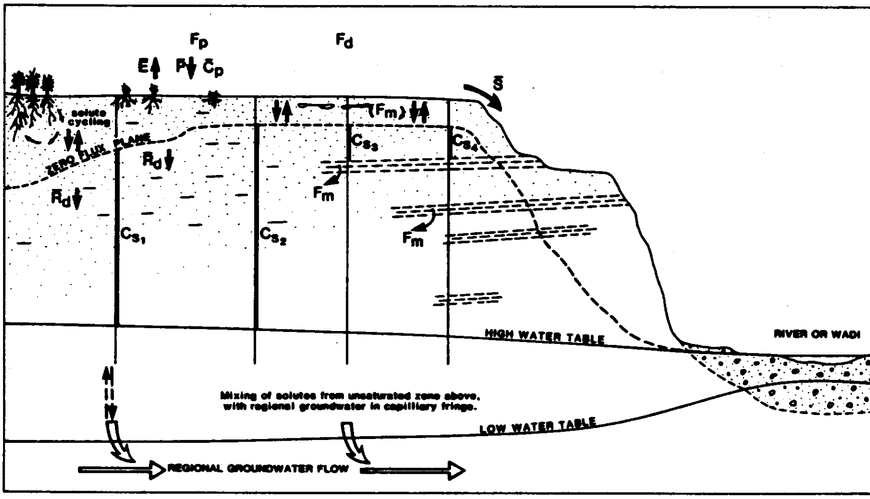

Fig. 20.1 Schematic representation of solute movement and recharge via the unsaturated zone. The symbols are defined in the text. Sections of unsaturated zone profiles $C_{s\,1-4}$ *useful for direct recharge estimation are shown in solid lines*

Nutrient cycling by plants may affect solute movement (including that of chloride) on an annual basis, but, in stable landscapes, the amounts removed annually by plant uptake are balanced by the amounts released by plant decomposition, i.e. a steady state should have been achieved. This assumes that there are no additions of the solute in fertilisers or permanent removal by crop harvesting (including the export of grazing animals).

The solute concentrations in the soil or in the upper unsaturated zone will vary seasonally or annually depending upon the intensity of the moisture flux due to the incident rainfall and evapotranspiration. Complex movement of solutes both upwards and downwards may take place in response to water movement, which in turn depend upon the prevailing water potential gradients. A "zero flux plane" (ZFP) exists (Wellings and Bell, 1980) which effectively separates moisture and solutes moving upwards (evapotranspiration) from that moving downwards (drainage). The position of the ZFP will shift seasonally between the surface and a depth of several metres and may, in some places and at certain times, be coincident with the water table in which case discharge may occur. Its position will also vary spatially in response to root development. However complex the soil moisture distribution might be in the soil zone, therefore, the transfer of moisture/solutes at depth will be a relatively straightforward process. Under conditions of recharge a maximum depth can thus be defined at which a net, steady state, moisture and solute transfer should take place towards the water table. The amount of solute crossing the ZFP would be expected to vary in relation to antecedent rainfall over one or more seasons and some oscillation in the solute profile would then occur. A detailed discussion of the transmission of solutes across the ZFP is given in Wellings and Bell (1980). The average composition of interstitial water in this profile (C_s) will, under steady state conditions, be proportional to the concentration factor, P/(P-E), assuming no loss of solute to minerals and that the water and "inert" solutes are transported at the same rate.

In the steady state, the water balance equation can therefore be given by:

$$R_d = P - E - S \tag{20.1}$$

where R_d is the direct recharge flux and S is the surface runoff flux and all quantities in eqns 20.1-20.6 are time- and space-averaged. Providing surface runoff is negligible ($S \approx 0$), this leads to:

$$R_d = P - E \tag{20.2}$$

Similarly the solute balance is given by:

$$F_p + F_d = F_s + F_m \tag{20.3}$$

where F_p and F_d are the average precipitation and net dry deposition fluxes (= input), respectively, and F_s and F_m are the net steady state output fluxes in the drainage water and the net flux of the solute precipitated or adsorbed by minerals (dissolution and desorption give a negative flux), respectively. F_s is given by the output water flux multiplied by the solute concentration (approximately averaged), i.e.:

$$F_s = R_d\ C_s \tag{20.4}$$

where C_s is the average concentration of the reference solute in the below-ZFP water. If we assume $F_m = 0$, then eqns 20.3 and 20.4 combine to give :

$$F_p + F_d = R_d\ C_s \tag{20.5}$$

or on rearranging:

$$(F_p + F_d)/C_s = (PC_p + F_d)/C_s \tag{20.6}$$

Hence the amount of direct recharge can be estimated from a knowledge of the volume-averaged concentration of the reference solute in the rainfall (C_p) and in the deep interstitial water (C_s), the long-term average annual precipitation (P), and the net dry deposition flux of the reference solute (F_d). Note that if $F_d = 0$, then the fraction of the rainfall contributing to direct recharge is simply given by the ratio C_p/C_s.

To recapitulate, the steady state model is subject to certain asumptions:

(1) since there is a time lag in solute input to the unsaturated zone and its output to the saturated zone, it must be assumed that no major climatic change has occurred over this period;

(2) that there have been no external, e.g. fertiliser, additions nor recent changes in atmospheric pollution;

(3) that there is no <u>net</u> change in storage of the "reference" solute above or below the ZFP, either by (a) plants or animals; or (b) by mineral precipitation/dissolution or adsorption/desorption. In the first instance this assumption should be valid if there have been no significant natural vegetation changes or changes in agricultural practices.

In principle, it may be possible to use as a reference any solute that is not released by weathering or removed by precipitation; even interacting solutes, e.g. cations on clay, in theory, may be able to be used since the quantity of exchangeable ions are often effectively constant and need not lead to a net change in storage. In practice the most inert solutes are likely to prove the most reliable and for this reason chloride has been used here for the recharge calculation although SO_4, NO_3 and SEC_{25} (specific electrical conductance at 25°C) have also been considered. It is possible that only a restricted portion of the unsaturated zone profile may be usable for recharge estimation (Fig. 20.1). For example, the presence of certain lithologies, e.g. residual marine bands, may release chloride. The possible development of solute marine bands is summarised in Fig. 20.2. Profiles 1 and 2 represent steady state drainage under different vegetation/soil conditions where evapotranspiration rates differ. Profiles 3 and 4 represent two possible cases where solute compositions have been modified by reaction and/or changes in storage. Only the upper part of profiles 3 and 4 would be of value in recharge calculations and, in certain reactive lithologies, no steady state profile may be developed at all.

The drainage compositions, C_{s1-2}, would be expected to be similar to those encountered at the surface of the water table. However the composition of the saturated flow will have been modified by the incoming lateral flow with higher salinity.

Therefore, water table samples taken from shallow wells are unlikely to be reliable for accurate recharge estimates. However, since chloride is not lost during drainage and saturated flow, the shallow groundwater chemistry, or river baseflow, can always be used in areas of active recharge to derive a minimum figure for total recharge (Eriksson, 1976) and could be more widely used in regional water balance studies.

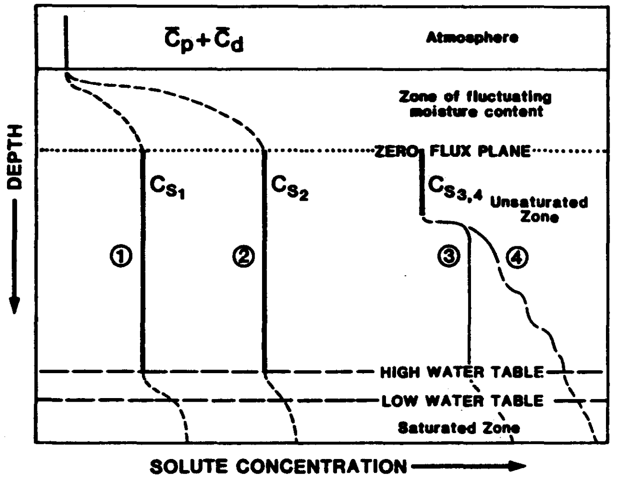

Fig. 20.2 Idealised solute concentrations developed during percolation in profiles C_{s1-4}. Steady-state sections are indicated by solid line

20.3 Methodology

Sampling of the unsaturated zone has so far been carried out using three different methods depending on the lithological nature of the terrain and the logistics. As the project has developed, the simpler techniques have proved to be the more effective.

Dry percussion drilling In Cyprus, profiles of up to 30 m were obtained by a wireline percusion technique, drilling dry, using a claycutter with a one-way valve. Metal casing was advanced every metre to avoid contamination. Samples were bulked every 20 cm and collected in polythene bags; drilling progress was typically 4 m per working day. This technique was attempted in Sudan but was unsuccessful due to the more indurated and clay-rich nature of the terrain, as well as logistical problems.

Power augering The principle technique used in Sudan involved using a Land Rover- mounted, petrol driven lightweight auger with 2-inch diameter flights. Samples were obtained to a maximum of 10 m by this method which proved the limit in this terrain (sandy clays overlying Cretaceous sandstone). The sampling method adopted involved the collection of comminuted sandy material delivered in sequence at the surface; each metre of material was then quartered and stored in polythene bags. Slight heating of samples occurred during drilling which might have caused some water vapour loss but this was not considered to be significant. Isotope measurements were made on these samples and no fractionation was observed.

Dug well sampling In order to take advantage of traditional construction methods, a dug well sampling programme was organised in Sudan. Initially, side wall sampling of a recently dug well at 25 cm intervals to depths of 30 m or greater was carried out using both a portable electric drill and an adze. However this technique proved to be operationally difficult and ran the risk of prior loss of moisture and capillary migration of solutes to the side wall. This technique was abandoned in favour of the "low technology" solution which is now being routinely used. Well-digging teams collect fresh material from the well at pre-determined intervals and/or at the end of each day's shift. Samples are collected in glass fruit bottling jars with rubber seals. Supplementary samples have also been obtained from shallow pits and excavations (up to 2 m).

Samples obtained by any of the above techniques have been used for soil moisture and solute determinations; however only percussion drilling and dug well sampling methods are suitable for isotopic determinations (including Tritium). Typically 50 g samples of moist disaggregated sediment are dispersed and elutriated in 30 ml aliquots of distilled, demineralised water for at least an hour, with occasional stirring; the optimum period is determined for each lithology to ensure effective elutriation. Supernatant solutions are then filtered (0.45 μm) prior to analysis.

Moisture content and SEC_{25} have generally been determined in the field and in this way, it is possible to make preliminary interpretation of soil moisture conditions and recharge estimations using SEC_{25}. Laboratory analysis was carried out for Cl and NO_3 using automated colometry; sulphate and cations were analysed if required by ICP-OES. Moisture was extracted for stable isotope determination by azeotropic distillation with toluene. In nearly all instances, agreement between duplicates was within ±2 ‰ for δ^2H and ± 0.2 ‰ for $\delta^{18}O$; duplicates falling outside this range were repeated until satisfactory. Isotopic analysis was carried out by conventional mass spectrometry, namely the reaction of 10μl water with heated Zn shot for δ^2H and the equilibration of 5 ml water with CO_2 of known isotopic composition for $\delta^{18}O$. Some Tritium analyses of sandy material were carried out as an independent check of the results from the solute profiles. Water samples were obtained by vacuum distillation.

20.4 Results

20.4.1 Cyprus

Recharge investigations were carried out on the Akrotiri peninsula in the extreme south of the island of Cyprus. Drilling was carried out in unconsolidated deposits of Recent age comprising mainly fine grained sands, where the unsaturated zone was at least 10 m thick. These studies were linked with a long-term lysimeter study of recharge (Kitching et al., 1980). Vegetation varied from grassland to forest; twelve boreholes were drilled in an area of 6 km^2 in locations representative of all land use types. The sites drilled

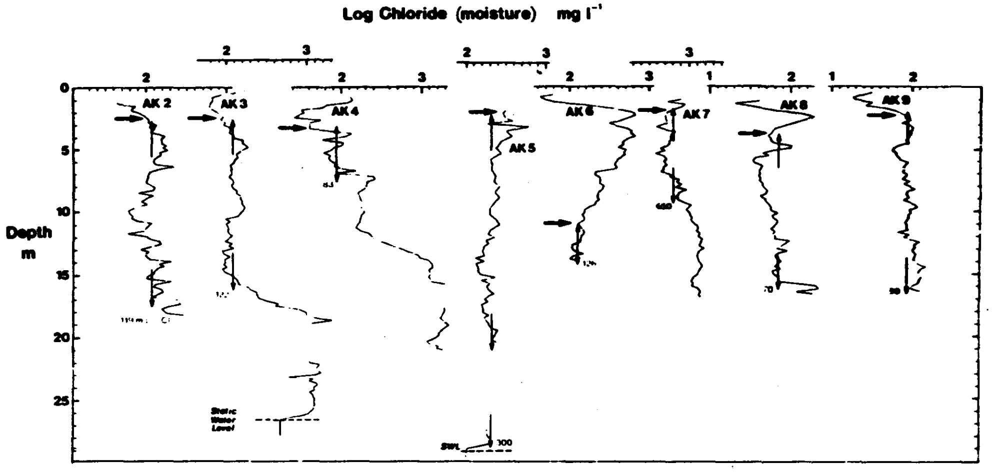

Fig. 20.3 Chloride profiles (log Cl) from eight boreholes drilled at Akrotiri, Cyprus

Vertical arrows indicate the profile length over which steady-state conditions are considered to exist. Horizontal arrows indicate the likely maximum depth of the zero flux plane (ZFP). Mean chloride concentrations for the intervals are given in mgl^{-1}.

represented undisturbed conditions with some animal grazing but with no inputs of artificial fertilisers. Of the twelve boreholes drilled, eight provided profiles to depths (15 m - 29 m) suitable for interpretation (Fig. 20.3), the others being abandoned due to drilling breakdowns of various kinds. A detailed discussion of the results is given in Kitching et al. (1980), Edmunds and Walton (1980) and Edmunds (1983), but a summary of the essential features is given here:

(1) Below a certain depth, chloride concentrations reach a relatively constant value. This depth (generally 2-3 m) is interpreted as the effective zero flux plane for solutes.

(2) At depth, a nearly constant interstitial water chloride concentration, C_s, is developed in all profiles with the possible exception of AK 7. There are fluctuations about this mean value but there are no significant trends with depth. Indeed, the amount of variation about the mean value gives an indication of the reliability that can be placed on the recharge estimate derived from a profile. The mean chloride concentrations also vary from profile to profile and this is thought to reflect genuine variation in the rate of recharge.

(3) Steady state conditions are found to persist to the water table (e.g. AK 5) or to the maximum depth

drilled (AK 8, 2, 9). In other boreholes there is an indication of chloride addition from the aquifer: in AK 3 there is an abrupt increase in chloride which coincides with a marine horizon (shelly gravels) and the profile below this cannot be used. In AK 4 there is a suggestion of continuous uptake of chloride through the succession, and only a limited portion of the profile has been used.

(4) Distinct peaks are apparent within the steady state sections of some profiles (e.g. AK 2), which indicate that homogenisation by dispersion does not take place. The peaks of AK 2 and possibly other profiles have an apparent periodicity of less than 1 m and are considered to represent cyclic, mainly annual, inputs of solutes from the soil zone (Edmunds and Walton, 1980).

Those sections of the solute (chloride) profiles that were used in recharge calculations are indicated in Fig. 20.3. They represent an integrated record of recharge over a period of time ranging from 1978 back to the late 1940's.

Table 20.1 Mean annual rainfall (3 year and 25 year) for stations at Akrotiri (Cyprus) with the average chloride deposition (wet and dry) over the three year period 1977-80

Year	Rainfall (mm)	Mean total deposition of chloride (kg/ha/y)
AKROTIRI (RAF)		
1977/78	455	55
1978/79	246	43
1979/80	506	77
AKROTIRI VILLAGE		
1977/78	426	66
1978/79	235	57
1979/80	564	96
MEAN VALUE	406 (25 y)	66 (3 y)

The 25 year mean annual rainfall value for Akrotiri (406 mm) was used for calculations and the 3 year mean chloride deposition flux including that of dry deposition is given in Table 20.1. The 3 year mean chloride deposition flux was used in conjunction with the 25 year rainfall value in the recharge calculations. There is some fluctuation in the chloride deposition both over the period of investigation and between the two stations. This is reflected in the total rainfall figures. The use of long term means in this technique is advantageous since the profile itself represents the average input over a number of years. The recharge estimates for the eight boreholes are given in Table 20.2.

At all eight locations, recharge was estimated to be in the range 10-94 mm/y. The variability is consistent with changes in topography and vegetation cover between sites with the highest values occurring in areas of sparse vegetation and low recharge (≈10 mm) taking place where bush vegetation exists. The results compare well with the indepedent recharge estimates obtained from the Tritium peak analysis.

Table 20.2 Data used from Fig. 20.3 for recharge estimation (Cl). Comparative results from Tritium profiles, (^3H), are given for reference

Borehole	AK 2	AK 3	AK 4	AK 5	AK 6	AK 7	AK 8	AK 9
Profile Interval(m)	2.4-17.3	2.4-16.1	3.2-6.9	2.0-28.3	10.9-14 0	1.7-9.4	3.7-16.0	2.0-16.5
Length (m)	14.8	13.7	3.7	26.3	3.1	7.7	12.3	14.3
C_s* (mg/l)	119	122	83	200	126	650	70	90
log C_s	2.076	2.086	1.920	2.301	2.100	2.813	1.845	1.954
R_d (Cl) (mm/y)	56	55	80	33	53	10	94	74
R_d (^{3}H) (mm/y)	52	53	N/A	22	N/A	N/A	62	75

* C_s is the mean interstitial water chloride concentration estimated by elutriation of samples from the profile intervals indicated in Fig. 20.3.

20.4.2 Sudan

Estimation of direct recharge was carried out as part of a wider study of the recent and palaeorecharge history and water resources of central Sudan (Darling et al., 1987). Profiles were obtained by augering and from dug wells at interfluve sites in the vicinity of Abu Delaig , 100 km ENE of Khartoum. Abu Delaig lies on a wadi system draining into Wadi el Hawad. Profiles were therefore obtained from indurated sediments including cemented sandstones in contrast to the unconsolidated deposits found in Cyprus.

This area is on the Sahel margin and renewable sources of groundwater are principally found in the Nile valley where there is groundwater recharge from the river Nile and along the wadi lines. Isotopic and geochemical studies demonstrate that much of the water being exploited from the deeper Nubian aquifer is palaeowater, recharged during the mid-Holocene (Darling et al., 1987). There appears to be little or no recharge via the wadi system to the deep regional aquifer at the present day.

Rainfall records are available for Abu Delaig from 1938 to the present day. Over the 29-year period 1938-1967, the mean annual rainfall at Abu Delaig was 225 mm but in the period 1971-1985 the corresponding value was only 154 mm. The period of study, 1982-1985, was the most arid on record with only about 15 mm recorded in both 1983 and 1984. Rainfall was sampled for chemical analysis during each of the years 1982-1985 and the mean concentrations of chloride are summarised in Table 20.3. These values represent the bulk deposition during the rainy season and it is assumed that during the dry season there is no input to or removal of chloride from the surface. A value of 5 mgl^{-1} Cl has therfore been adopted for C_p, although, as discussed below, this may not be valid as a long-term average.

Table 20.3 Chloride concentrations of rainfall (mean annual average) at Abu Delaig, Sudan, for the period 1982-1985 (mgl^{-1})

Year	Total rainfall	Mean Cl (weighted average)
1982	192	6.1
1983	15.5	10.6
1984	15.2	5.6
1985	118	2.4

Volume weighted 4 year average: 4.9 mgl^{-1}

Four solute profiles from the unsaturated zone at Abu Delaig illustrate the results for Sudan (Fig. 20.4) which are described in detail elsewhere (Edmunds et al., in preparation); hole A is a hand dug well and G, M, Q are auger holes. All profiles were obtained in November 1983 some three months after the very poor rainy season.

The moisture contents, plotted on a dry weight basis, range from 2% to 11% and strongly depend on lithology, with the highest values occurring in clay-rich horizons. The overall solute distribution is given by SEC_{25} which reflects the total mineralisation of the moisture profile. The chloride and nitrate profiles are also illustrated. Results are expressed as elutriate concentrations and then as calculated pore water concentrations after correcting for the dilution during elutriation. After correction, it was found that estimated pore water SEC_{25} values varied from 2190 to 79430 $\mu S\ cm^{-1}$, equivalent to a total mineralisation from about 1300 mgl^{-1} to values rather above sea water concentrations. These corrected SEC_{25} values should only be interpreted in a semi-quantitative sense because of the lack of strict proportionality between SEC_{25} and concentration. Nevertheless, they give a good indication of the general trend in mineralisation of the interstitial water down a profile.

Chloride concentrations generally mirror SEC_{25} indicating that Cl is an important constituent of the total mineralisation; sulphate is also present as a major solute, particularly in the upper section of the profiles (Edmunds et al., in prep). Nitrate is also an important constituent of profiles G, M and Q on the southern side of the wadi with highest concentrations up to 1000 mgl^{-1} NO_3-N in pore solutions. This is in contrast to hole A on the northern side of the wadi where background levels closer to 10 mgl^{-1} NO_3-N are found. The variation in nitrate concentrations is not yet fully explained but is likely to relate to different vegetation or cultivation on either side of the wadi; it is unlikely therefore that in this location nitrate could form a suitable reference solute for recharge estimation.

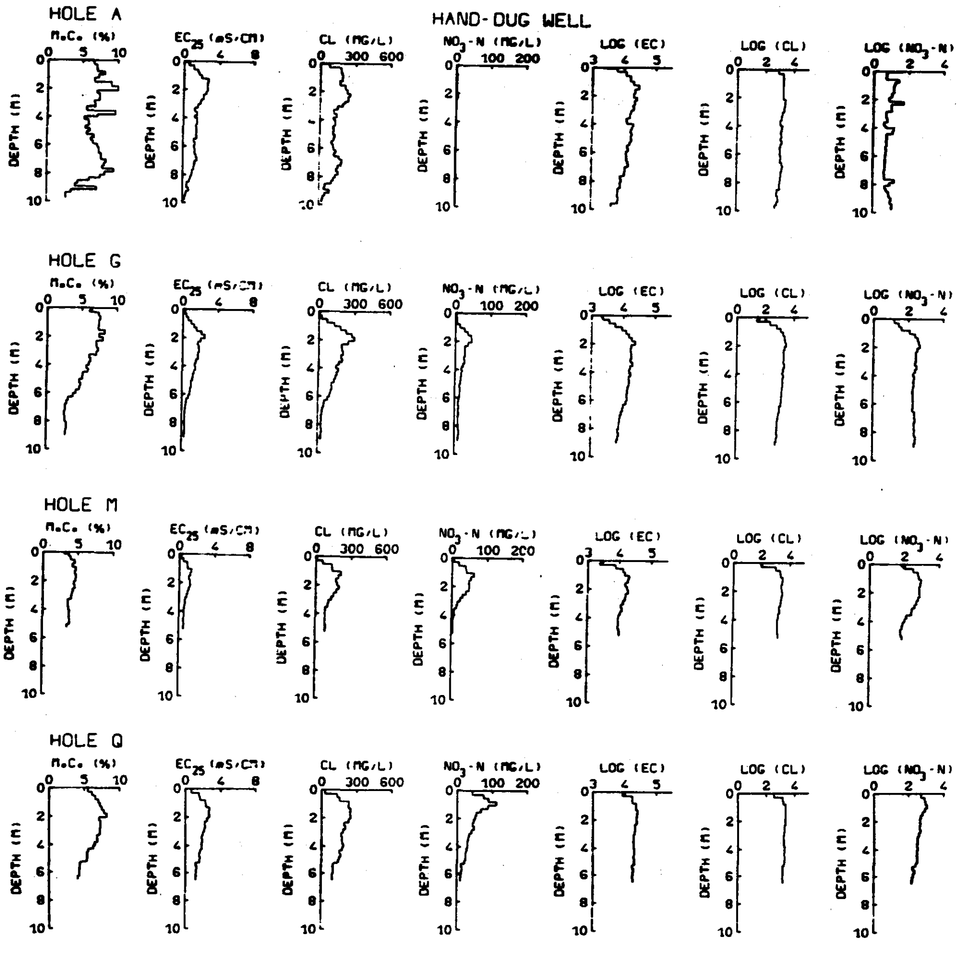

Fig. 20.4 Moisture and solute profiles for one dugwell (A) and three augers at Abu Delaig, Sudan

Data shown are: moisture content (M.C.) on dry weight basis; specific electrical conductance (EC_{25}) of the elutriate; chloride concentration of the elutriate (Cl); nitrate concentration for the elutriate (NO_3-N); log of specific electrical conductance of soil moisture after correction for moisture content (LOG EC); log chloride concentration of the soil moisture after correction for moisture content (LOG CL); log concentration of nitrate in soil moisture after correction for moisture content (LOG NO_3).

Chloride concentrations increase with depth over the first 1-2 m and below this reach near constant values. They may be interpreted as typical recharge profiles with a zone of fluctuating moisture content above a zero flux plane at about 2 m depth. The absence of any salt accumulation indicates a net recharge for these profiles and a steady-state portion of several metres can be observed in each profile. None of these profiles reaches the water table but data are available from several dug wells in the vicinity where much lower chloride concentrations are observed (in the range 40-60 mg Cl l^{-1}).

Table 20.4 Mean chloride concentrations in interstitial waters from profiles at Abu Delaig (Sudan) and the corresponding estimates of direct recharge

Profile	Interval (m)	Mean Chloride (mgl^{-1})	No. of Meas.	Rd (mm)
A	1.0 - 9.80	1288	39	0.78
B	1.05 - 2.85	1738	8	0.58
C	1.05 - 5.00	2239	20	0.45
D	2.24 - 4.20	2228	8	0.45
E	0.80 - 6.95	783	25	1.28
G	2.05 - 8.80	1675	27	0.60
H	0.80 - 3.30	1208	10	0.83
J	0.15 - 2.80	1120	8	0.89
K	0.80 - 2.30	971	6	1.03
M	0.80 - 5.30	1936	18	0.52
N	1.55 - 1.80	3936	22	0.25
P	0.55 - 3.00	173	10	5.78
Q	0.80 - 6.55	1845	23	0.54
S	1.30 - 4.30	1663	9	0.60

Mean chloride concentrations were obtained after correction for the dilution resulting from elutriation.

The mean chloride concentrations have been calculated for the steady-state sections of all 14 profiles taken from an area of some 6 km^2 of interfluve terrain (Table 20.4). The interstitial chloride concentrations range from 1000 to nearly 4000 mgl^{-1} and there is some correspondence between concentration and site location. The highest value (N) occurs in a profile taken from a sandstone ridge above the main wadi where some run-off might have occurred. The lowest value (K) is from a relatively clay-rich profile from flat lying ground. Using steady-state interstitial water chloride concentrations (C_s), a mean chloride concentration in rainfall (C_p) of 5 mgl^{-1} and the long-term average rainfall (P), values for R_d are calculated. Excluding the wadi profile, these lie in the range 0.25-1.28 mm/y. A regional long-term average (R_d) of 0.72 mm/y may be calculated for the direct component of recharge via interfluve areas in this area of Sudan.

20.5 Discussion

A main advantage of the chloride profile technique is that results are derived in part from long-term data, compared with direct measurements of flow in the unsaturated zone where results are typically obtained on an annual basis. The solutes in the unsaturated zone act as a totalising rain gauge with storage of moisture over decades or centuries. Profiles therefore can act as a means of obtaining long-term averages of recharge and also as a means of detecting climatic change (Edmunds and Walton, 1980; Allison et al., 1985). In exceptional circumstances, where the data are not smoothed by the dispersion it may be possible to discern annual increments in the profile. This can be seen in the case of the chloride profile AK 2 (Fig. 20.3) where individual annual peaks are visible (Edmunds and Walton, 1980); using the SEC_{25} profile from the same borehole gives a good correlation with the oscillations in rainfall since the 1940's.

The chloride profiles from Sudan, indicating much lower net recharge rates, indicate that water is stored in the unsaturated zone for much longer periods. Using a value for the porosity of 30%, saturation at 25% and a recharge rate of 1 mm/y, a 25 m profile can be storing water which has been in transit for about 2000 years. The shape of some profiles from Sudan indicates that slightly less saline water is found with increasing depth which may be a reflection of slightly higher recharge rates in earlier centuries. An alternative explanation could be that there is a component of upward water vapour in these profiles and this is suggested by isotopic results ($\delta^{18}O$, $\delta^{2}H$) from some of the profiles in Sudan (Darling et al., 1987). In particular, the isotopic composition of the deepest moisture is similar to that of lower salinity water contained in the saturated zone. However, although discharge (loss of water vapour) may be occurring during certain periods <u>net</u> recharge is indicated on the basis of the chloride profiles. The profiles might therefore represent the complex result of both processes, with recharge only activated during episodes of higher rainfall. Resolution of this problem requires further data from similar areas especially from those with higher mean annual rainfall.

Just as important as measuring and obtaining a reliable figure for the mean interstitial water chloride concentration, is the reliable estimation of atmospheric inputs. This is particularly difficult in most semi-arid regions where long-term records of the chloride composition of rainfall and dry deposition are practicaly non-existent. In the present study, reliable data have been obtained for total solute deposition over periods up to 4 years, and as indicated, the rainfall in Sudan during this study was well below the long-term average. Longer runs of data would improve the reliability of the recharge estimates. There is also the problem that rainfall samples can contain dust and that both rain and dry deposition are therfore likely to be sampled together during the rainy season. The effective rainfall chemistry used for recharge estimate purposes in Sudan is rather higher in chloride than that expected from its continental position. In this study it has been assumed that, during the rainy season all chloride is washed into the profile, and that dust is "on the move" during the dry season; thus net deposition is insignificant except during the rainfall episodes when sampling of bulk deposition took place.

20.6 Acknowledgements

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21 RAINFALL - RUNOFF - RECHARGE RELATIONSHIPS IN THE BASEMENT ROCKS OF ZIMBABWE

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21.1 Introduction

As a result of the recent drought in Central Africa an accelerated programme of drought relief was commissioned by the Government of Zimbabwe and the European Economic Community in Victoria Province, Zimbabwe (see Fig. 21.1). Over an area of 22000 km², 282 boreholes were sited, drilled and equipped with handpumps. The area is one of variable relief on the edge of the Zimbabwe plateau and extends down toward the river Limpopo valley. The area is underlain by Basement granites and gneisses. The principal aquifer is composite composed of weathered regolith of low permeability, high storage, overlying fissured bedrock of high permeability and low storage.

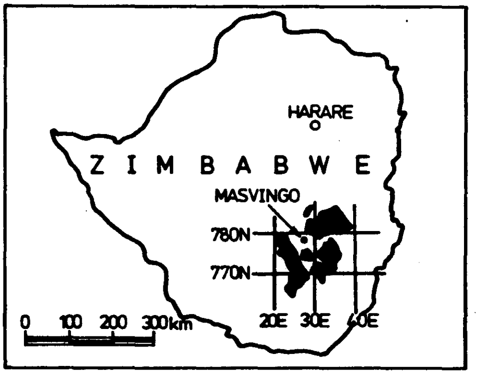

Fig. 21.1 Project location map, Masavingo, Zimbabwe

No borehole hydrographs are available to indicate that recharge to Basement aquifers is taking place in the project area. However, surface water hydrographs show considerable baseflow from aquifers in every year for which records are available. This is strong evidence for recharge. Hydrochemical evidence also indicates the probability of active recharge. Furthermore, by analogy with other areas in central Africa (Foster et al., 1982; Houston, 1982) it would appear likely that recharge will occur whenever the annual rainfall exceeds a threshold value. Lateral recharge from other aquifers or surface water is also a possibility, but is likely to be on a localised scale and is not considered further here.

21.2 Hydrometeorology

Within the project area there are 22 rainfall stations with daily data in one case dating back as far as 1900. Only three stations have class A pans for evaporation and records only go back as far as 1964. No systematic data is available for the calculation of evapotranspiration by an energy balance method such as Penman's so that it has been necessary to convert pan evaporation to evapotranspiration using monthly conversion factors determined empiricallly elsewhere in Africa (Aune, 1970; Heederick et al., 1984; Pike, 1971; Riou, 1984).

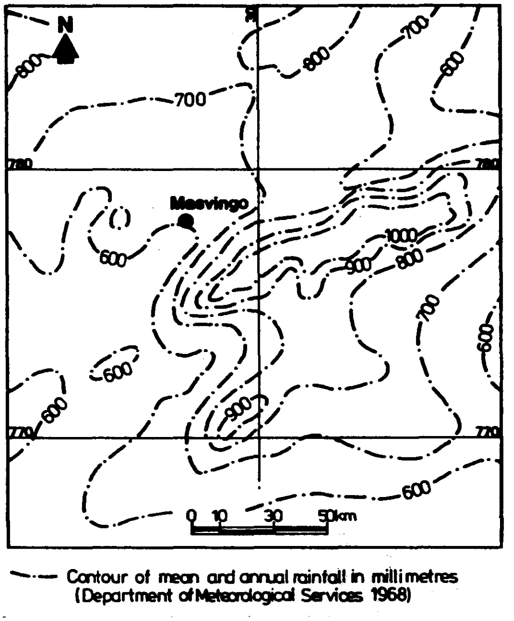

Fig. 21.2 Mean annual rainfall for Victoria Province

The map of mean annual rainfall produced by the Department of Meteorological Services in Zimbabwe is summarised in Fig. 21.2. It is clear that rainfall is orographically controlled and dependent upon the south-east Trade winds bringing moist air up the Limpopo valley from the Indian Ocean during summer. Summer lasts from October to March when the Inter-Tropical Convergence zone moves south over northern Zimbabwe.

Changes in diurnal and seasonal rainfall are shown in Fig. 21.3 for Masvingo over a thirty year period. The rainy season usually starts in the second half of October and lasts for five months to the end of March. The most reliable months for rainfall are November, December and January. The daily intensity is also greatest during these months and greatest during the afternoon. This means that the bulk of the rainfall is concentrated in a relatively short time span,

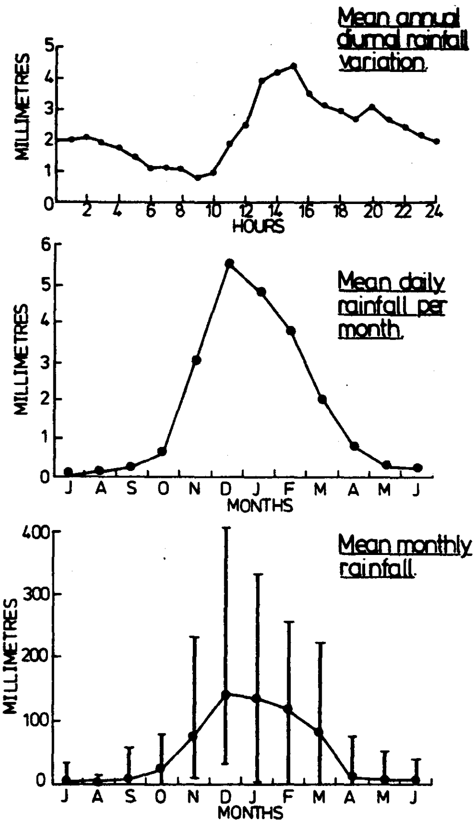

Fig. 21.3 Diurnal and seasonal changes in rainfall at Masavingo for the years 1941 to 1971

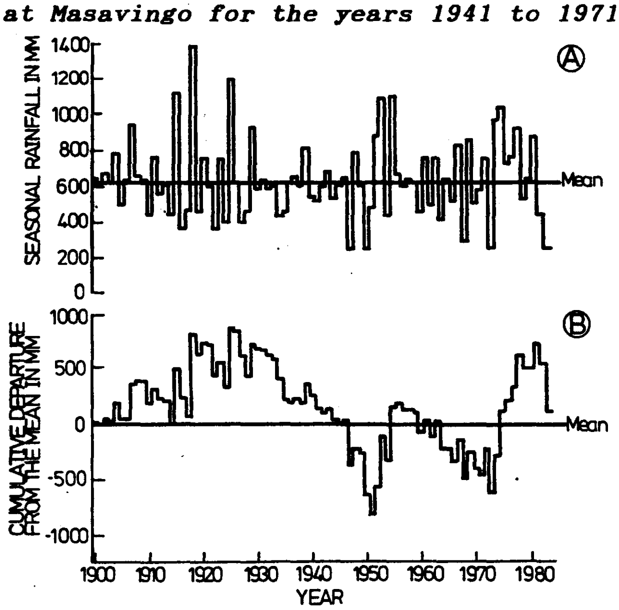

Fig. 21.4 Annual rainfall time series for Masavingo

enhancing the potential for both runoff and recharge to aquifers but limiting it to short periods of the year.

Longterm variations in rainfall are also of importance to recharge estimates. The mean seasonal rainfall at Masavingo over 84 years since 1900 is 623 mm but there is variation between 40% and 220% of this figure. The mean annual rainfall for the project area as a whole, found by integrating the areas on the rainfall contour map is 728 mm.

The variation in annual rainfall is shown in Fig. 21.4. The periods 1905-25, 1951-55 and 1970-80 were wetter than normal whilst the periods 1925-35, 1945-50 and 1965-70 were drier than normal. The intensity of any dry period can be determined from the steepness of the descent of the graph in Fig. 21.4b. The recent drought intensity has never been matched since records began but was almost matched in 1949-51. However the antecedent rainfall was much less in both 1949-51 and 1968-73 than in 1981-84. Assuming that recharge to Basement aquifers is largely precipitation dependent, this suggests that although groundwater levels may be falling sharply now they may not be as low as they were around 1950 and 1970.

Over the entire period there is no evidence for a trend toward drier conditions. However, there is some evidence to suggest poorly developed periodicities or rainfall cycles. The evidence comes from the autocorrelation and power spectrum of the annual rainfall series (see Fig. 21.5). The most significant periods are of 23 years (with a harmonic at 11.6 years) and a shorter cycle of 14 years (with harmonics close to 7 and 3.5 years). Both of these periodicities are commonly found elsewhere in Africa (Houston, 1982; Lamb, 1982) and India (Vines, 1986). A reconstruction of the principal periodicities suggests that low rainfalls may continue to be a feature through the 1980's with a return to generally wetter conditions by the mid 1990's.

Mean monthly values of pan evaporation and potential evapotranspiration are shown in Fig. 21.6 for a ten year period. Evapotranspiration reaches a peak during the rainy season but the potential for excess rainfall contributing to runoff or recharge is clearly seen, despite the fact that on an annual basis evapotranspiration amounts to an average 1246 mm.

21.3 Baseflow analysis

Within Victoria province four gauging stations are available with average daily flow data for periods varying from 8 to 17 years. Catchment areas vary from 100 to 5390 km^2, all largely on granitic and gneissic rocks.

The baseflow component of the river hydrographs shows up clearly on both arithmetic and semi-logarithmic plots (see Fig. 21.7).

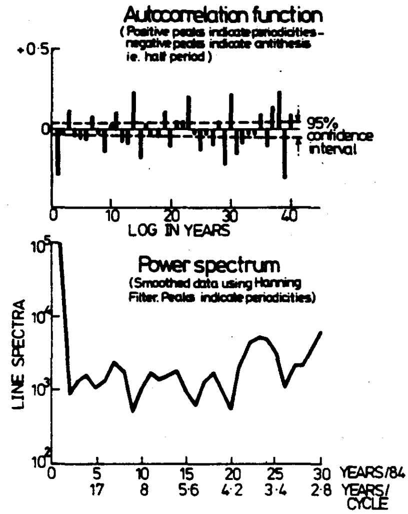

Fig. 21.5 Time series analysis of annual (seasonal) rainfall at Masavingo for the years 1900 to 1983

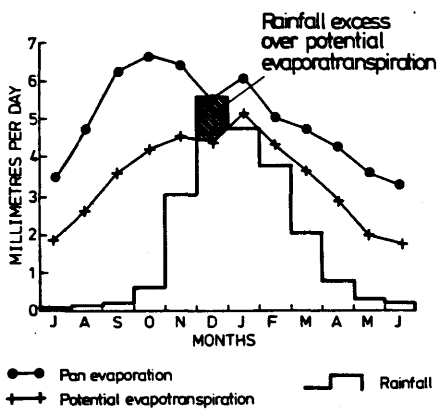

Fig. 21.6 Mean monthly values of evapotranspiration and rainfall at Masavingo for the years 1974 to 1983

Baseflow discharge represents water which has percolated through the soil and vadose zones to the groundwater table but which subsequently discharges to rivers by means of gravity drainage. The simplest water balance for a catchment is given by:

discharge = recharge + change in storage

Thus over periods of equivalent storage, recharge to groundwater is given by the baseflow component of the hydrograph. Each component of the hydrograph can be approximated by a relationship of the form

$$Q_o = Q_t K^t$$

as shown by Barnes (1939), where Q_o and Q_t are the discharges at the beginning of the measurement period and after time t respectively and K is the recession constant. In the case of baseflow, K is a function of the aquifer transmissivity, storage and catchment geometry.

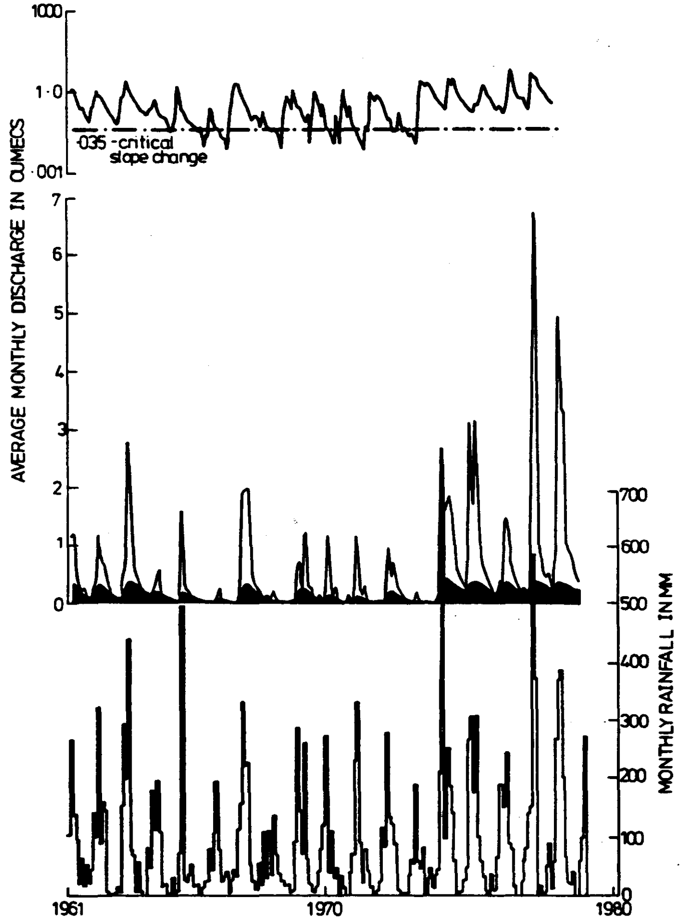

Fig. 21.7 The relationship between rainfall, runoff and baseflow for the Mzero catchment

All hydrographs examined exhibited a critical change of slòpe at extreme low flows, with marked increase in the rate of recession. The critical change in slope represents an increase in the rate of depletion of active storage in the catchment. Since it always occurs at the same discharge level for each catchment it also occurs at the same groundwater head distribution each time. It is considered likely that at this critical point an upper zone becomes depleted and a lower zone becomes active with different aquifer characteristics. This lower zone would have higher permeabilities and the critical level must thus represent the change from regolith aquifer discharge to underlying fissured bedrock aquifer discharge. It is also important to notice that baseflow continues thoughout the dry season in all catchments in five out of six years, pointing to the importance of the regolith as a storage unit.

A regression analysis between the baseflow and rainfall was carried out for each catchment. The slope of the regression line represents baseflow as a proportion of rainfall and the intercept of the regression line with the rainfall axis at the point where baseflow is zero gives the minimum critical rainfall for baseflow to have occurred (see Table 21.1).

Table 21.1 Catchment characteristics for four stations

Station	Chiredzi	Mzero	Musokwesi	Lundi
Catchment area km²	1029	100	246	5390
Mean rainfall mm	818	1055	1126	797
Mean baseflow mm	31	51	32	6
Baseflow as % rainfall	4.7	8.5	4.3	1.9
Minimum rainfall for recharge mm	157	452	379	463
Recession constant	.990	.994	.991	.996
Period of record	66-77	61-78	69-78	70-78

21.4 Hydrochemical analysis

The variation in electrical conductivity of groundwaters is shown in Fig. 21.8. Regolith groundwaters display a tendency for values less than 150 mmhos/cm to be associated with high rainfall and runoff areas. All major ions show a similar pattern as a result of their general linear relationship with electrical conductance.

In the case of chloride, the principal source in groundwaters is from the atmosphere since there are no evaporite sources and there is unlikely to be any large contribution from the granite and gneiss host rocks. Based on this assumption the ratio of chloride in rainfall to that in groundwater is proportional to recharge thus:

$$\text{recharge mm} = \text{rainfall mm} \times \frac{\text{rainfall Cl mg/l}}{\text{groundwater Cl mg/l}}$$

The relationship between chloride and rainfall is shown in Fig. 21.9. There is some suggestion here that groundwaters from areas of gneiss tend to have a higher chloride content. For each rock type a significant relationship exists between rainfall and chloride content strongly suggesting that recharge is a function of rainfall.

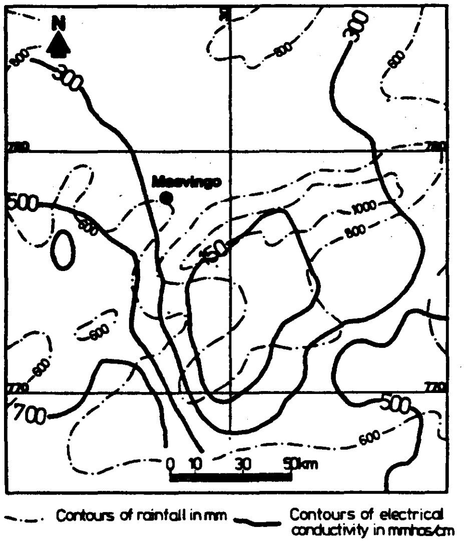

Fig. 21.8 The electrical conductivity of regolith groundwaters in comparison with mean annual rainfall

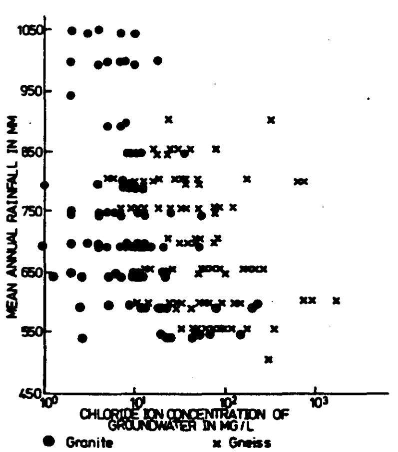

Fig. 21.9 The relationship between chloride ion concentration in groundwater and rainfall at the same sample site. Rainfall interpolated from isopleth map

The chloride content of rainfall in Zimbabwe is not known but by analogy with similar areas might be expected to be about 0.5 mg l^{-1} (Foster et al., 1982). Using this value and rainfall values for the area, estimates of recharge can be made (see Table 21.2). The results show that for granite aquifers recharge is likely to vary from 2-3% rainfall. For gneiss aquifers recharge rates of only 0.5-1% of rainfall are calculated. This may suggest another source of chloride in gneiss aquifers.

Table 21.2 Mean chloride content and recharge as a percentage of rainfall for different geological and rainfall environments

Rainfall (mm)	Granite Cl (mgl^{-1})	Granite recharge (%)	Gneiss Cl (mgl^{-1})	Gneiss recharge (%)
500	28	1.8	104	0.5
728	15	3.3	59	0.9

21.5 Recharge-runoff simulation

In order to assess recharge more accurately a simulation model was used which requires rainfall and evapotranspiration data as input and produces estimates of recharge and runoff as output. It is essentially a routing model for flow through the soil and unsaturated (vadose) zones to the water table and to runoff. At its heart is the calculation of soil moisture deficits by the method of Penman (1949) and Grindley (1967) which in their simplest form are given by:

$$SMD_{t+1} = SMD_t + E_t - RF_t \qquad \text{for } SMD_t > 0$$

$$SMD_{t+1} = E_t - RF_t \qquad \text{for } SMD_t < 0$$

where SMD_t is the soil moisture deficit at time t, E is the evapotranspiration component and RF is rainfall. The model requires the estimation of a root constant which operates a negative feedback mechanism by limiting evapotranspiration when soil moisture deficits are high. Any rainfall in excess of evapotranspiration and soil moisture requirements is routed downwards to the vadose zone subject to the infiltration capacity where a fraction becomes interflow runoff and the remainder recharges groundwater.

A great variety of such models have been produced for temperate climates where the results have been found to be generally good (Calder et al., 1983) but tend to underestimate recharge when used on a monthly timestep basis (Howard and LLoyd, 1979). Houston (1982) has shown that such models are equally applicable to dry tropical climates providing reasonable assumptions are made. Furthermore, the error in using monthly time steps as opposed to daily is less than other potential errors, for instance that involved in determining evaporation.

Using monthly rainfall and evapotranspiration data obtained as previously described, a simulation model was developed and calibrated on the period 1974 to 1983, a period when the average rainfall was 727 mm, close to the overall mean for Victoria province. A flow diagram for the simulation model is shown in Fig. 21.10 for run number 11. Runs 1 to 10 were used to test the sensitivity of the model by varying such factors as the infiltration capacity, the fractions of direct recharge and interflow, the maximum volumes held in surface detention and soil moisture storage, and the root constant (finally fixed at 100 mm). All values employed are considered to be realistic in the circumstances operating in Victoria province.

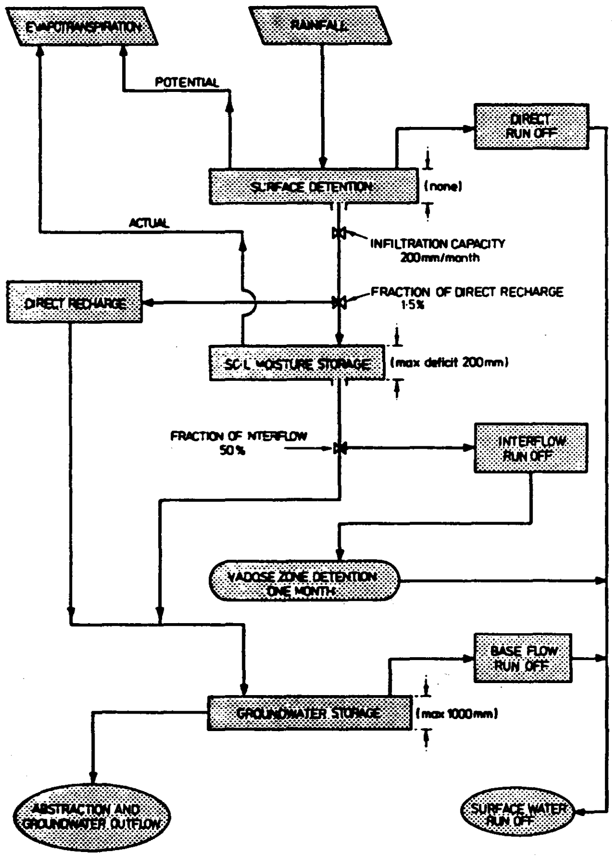

Fig. 21.10 Flow diagram for recharge-runoff simulation model

The output was checked against the Lundi river hydrograph as shown in Fig. 21.11. In general the results are good confirming the validity of the model. However, in detail the actual data from the Lundi river suggests that there is slightly more interflow and baseflow than the model predicts. This would have the effect of reducing recharge to groundwater marginally. The results of the model (given in Table 21.3) suggest that some errors may be present in the first year of simulation. Since the runoff is underestimated recharge is overestimated. After the first year the results settle down and mean recharge over an eight year period amounts to 22 mm equivalent to 2.5% of rainfall.

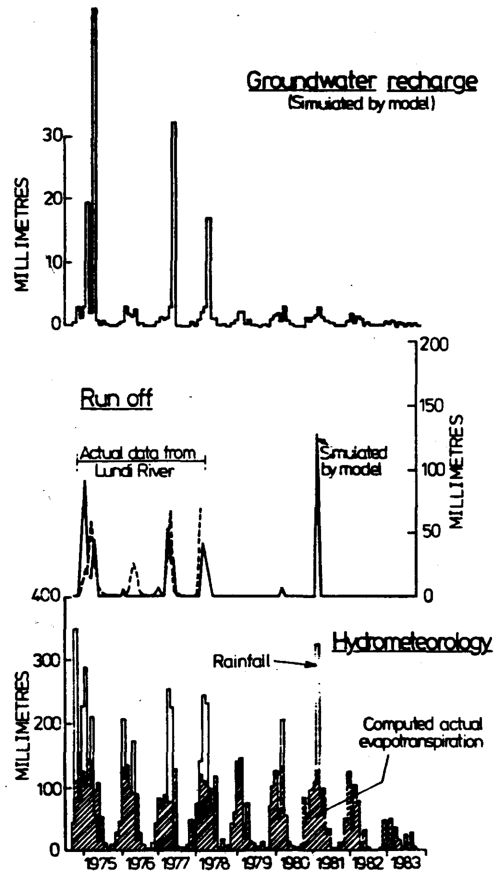

Fig. 21.11 Results of recharge-runoff simulation model

Table 21.3 Results of recharge-runoff simulation model for an idealised basement aquifer in Victoria Province

Year	1975	1976	1977	1978	1979	1980	1981	1982	1983
Rainfall mm	1511	720	763	919	521	634	905	441	211
Recharge mm	82	11	39	27	8	9	12	7	3
Recharge as % of rainfall	5.4	1.5	5.1	2.9	1.5	1.4	1.3	1.6	1.4

21.6 Recharge to basement aquifers

It is possible to compare the results of recharge estimated by three completely independent methods. In order to do this the results of each estimate have been plotted as recharge against rainfall in Fig. 21.12. From this it can be seen that the results compare very well. Over much of Victoria province recharge amounts to 2-5% of the rainfall on a long term basis. On a Province-wide basis with mean annual rainfall of 728 mm, recharge amounts to 12 mm/annum equivalent to 12000 m^3/annum/km^2.

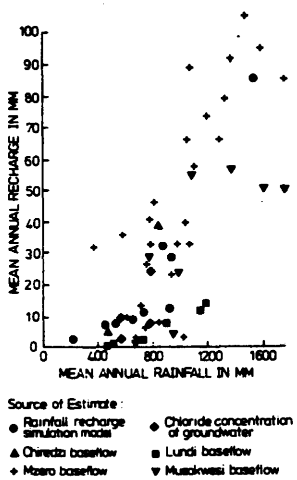

Fig. 21.12 Relationship between recharge and rainfall

In detail however, there is considerable variation in recharge resulting largely from the uneven distribution of rainfall both in space and time. On a spatial basis, those areas above 700 mm rainfall are more likely to receive significant recharge than those below 700 mm. Also, below 400 mm it is questionable whether recharge ever occurs. On a temporal basis recharge tends to be correlated with rainfall such that recharge is limited to short periods within the year and a few years with heavy rainfall are more important in producing recharge than many years of average rainfall.

If a conservative estimate of recharge is assumed to be 2% of rainfall in an area of relatively low rainfall which, for Victoria Province would be about 500 mm/annum, then it can be seen that recharge amounts to 10 mm/annum. For a rural development scheme the average demand of a handpump is 6500 m^3/annum assuming 0.5 l/s over a 10 hour pumping day, 360 day year. If it is assumed that the yield is totally reliant on precipitation recharge then an area of 0.65 km^2 is required for its sustenance at the recharge rate of 10 mm/annum. This is equivalent to a radius of 450 m around the borehole. Based on equilibrium conditions and average aquifer charcteristics for Basement rocks, the cone of depression due to a handpump would extend to at least 1000 m around the borehole and therefore recharge is not likely to be a limiting factor in resource development except in special circumstances.

21.7 Discussion

It is quite clear that in any assessment of recharge in similar environments it will never be adequate to assess recharge based on a short data length. Since the necessary data for a full recharge evaluation is rarely available it is also important to use a variety of techniques for cross reference and to investigate more closely that aspect of the hydrological cycle, namely rainfall, which is more frequently better documented and can lead to a greater insight into potential recharge patterns.

Reliable estimates of recharge to aqifers in central Africa are not common, but it is worthwhile briefly considering two other estimates. In an area on the edge of the Kalahari in Botswana, Foster et al. (1982) suggest that precipitation recharge is unlikely when the mean annual rainfall falls below 450 mm and the sand cover is greater than 4 m. In a much wetter area (mean annual rainfall 937 mm) of Zambia, Houston (1982) showed that recharge was not only dependent upon rainfall but also upon the vegetation type. Thus in areas of natural forest where evapotranspiration was high, recharge amounted to 80 mm or 9% of rainfall, whereas those areas of cleared ground with crops produced 281 mm of recharge or 30% of rainfall. In both areas of Zambia infiltration rates were extremely high and runoff negligible.

The conclusion drawn from this is that for large areas of Africa, recharge is dependent on rainfall both spatially and temporally. It is tentatively considered that rainfall below 400 mm is unlikely to produce recharge and above this value the amount of recharge will depend upon factors such as soil and vegetation type.

21.8 Acknowledgements

I would like to thank the Ministry of Water Resources and Development and Energy and the European Economic Community for allowing the results of this project to be presented here.

22 TWO GROUNDWATER MODELS FOR ESTIMATION OF NATURAL DIRECT GROUNDWATER RECHARGE - A COMPARATIVE STUDY IN SANDY TILL

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22.1 Description of the study area

The study was performed in moraine terrain close to Emmaboda in southeastern Sweden (Fig. 22.1). The sandy till covering the area is on average 4 m deep and underlain by granitic bedrock. The area is situated in the catchment area of the River Lyckebyan. The aquifers in the sandy till are small but still important as water supplies for small villages, farms and single households.

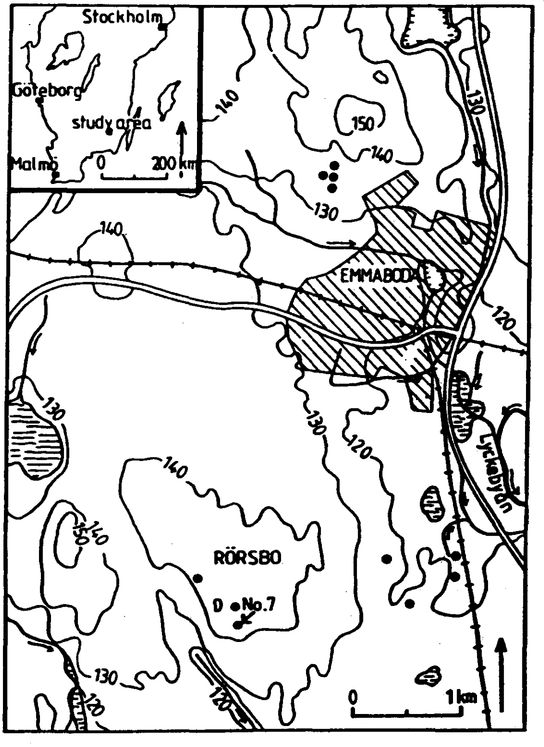


Fig. 22.1 Map of the study area, SE Sweden (dot = observation tube)

Groundwater levels have a yearly maximum in spring after snowmelt and a minimum in autumn. The groundwater is shallow. The yearly average groundwater level in the 11 observation tubes, run by the Geological Survey of Sweden (SGU), is approximately 1 m below ground surface and the yearly amplitude about 2 m. The groundwater level fluctuations are well correlated in time. A factor analysis already gave an explained relative variance of 93% by the first variable. However the analysis also indicated considerably greater fluctuations in the other tubes.

The investigations were concentrated on the immediate vicinity of observation tube No. 7 within the Rörsbo subarea. The groundwater level fluctuations at Rörsbo are extremely well correlated and could be said to be representative for the subarea. Observation tube No. 7 is located on a gentle slope close to a water divide and between a small grassland and a deciduous forest.

Close to the tube a 2.0 m deep pit was dug in August, 1984. A plough layer extended down to 20 cm and was underlain by a B-horizon down to 40 cm. The C-horizon was divided into two sub-horizons (40-100, 100- cm), where gravel and stones were more frequent at depth. Four vertical, parallel core samples (length: 5-10 cm, diam.: 7.2 cm) were taken at three different depths, for analyses of texture, water retention properties and saturated hydraulic conductivity in the laboratory (Figs. 22.2, 22.3). Despite a slow successive saturation from below during a period of 14 days, it was impossible to saturate the samples completely. The water constant at saturation was 4.4-5.7% lower than the corresponding porosity values calculated from the dry bulk and solid densities. The groundwater table was 1.5 m below the ground surface at the time of sampling.

Precipitation and snow depth are measured by the Swedish Meteorological and Hydrological Institute (SMHI) at Rörsbo. The other climatic data needed in the modelling (air temperature, vapour pressure, wind speed and cloudiness) were taken from SMHI's station in Växjö, 50 km NW. Average monthly precipitation and air temperatures are shown in Table 22.1.

Table 22.1 Average monthly precipitation (uncorrected) and air temperatures from the stations Rörsbo and Växjö respectively, run by SMHI (based on data from 1968-1985 and 1951-1980 respectively)

	J	F	M	A	M	J	J	A	S	O	N	D	Year
P(mm)	42	27	35	34	44	47	64	45	54	50	59	48	546
T(°C)	-2.8	-3.0	0.4	4.6	10.3	14.7	15.8	15.2	11.3	7.1	2.6	-0.8	6.4

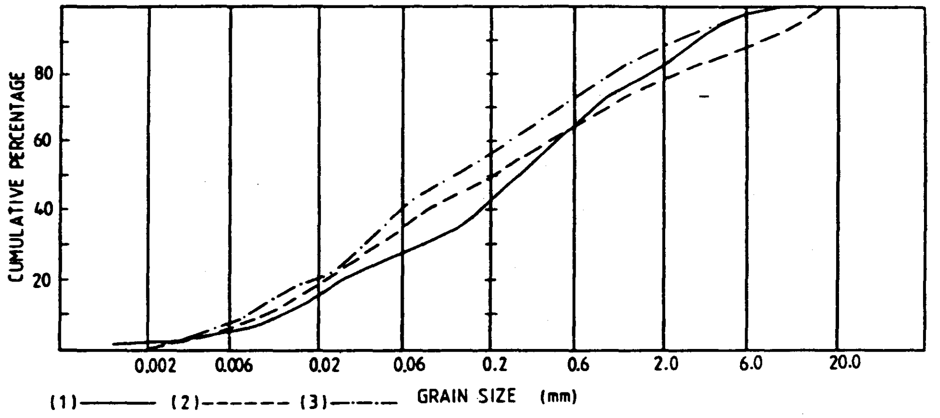


Fig. 22.2 Grain size distribution for the soil profile at SGU's obs. tube no. 7, Emmaboda (1) = 0-40 cm, (2) = 40-100 cm, (3) = 100 cm

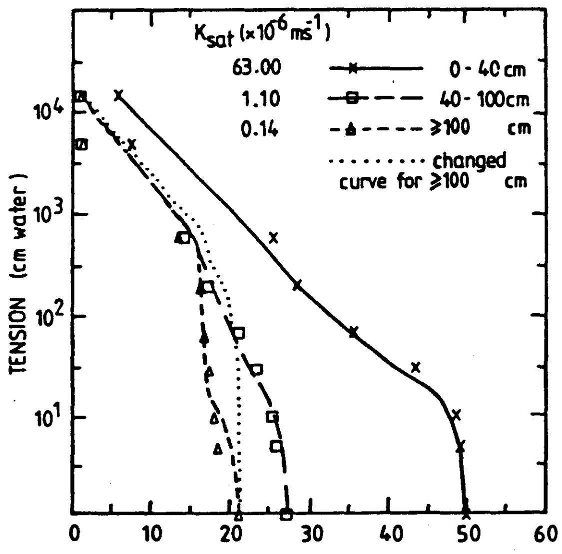


Fig. 22.3 Measured pF-values and fitted curves for the different layers (for the changed curve, see text)

22.2 Methods

22.2.1 Estimation of groundwater recharge from one-dimensional soil water modelling

A soil water flow model, with appropriate submodels for the boundary conditions, calculated groundwater recharge from inflow data. The model used was a one-dimensional soil water and heat model developed by Jansson and Halldin (1979). A detailed description of the model was given by Jansson and Halldin (1980) and only some main features will be given here together with adaptations to the studied problem.

The soil water flow, expressed in the units employed in this study, reads:

$$\delta\theta/\delta t = -\delta/\delta z\ [k(\delta S/\delta z + 1)] - s \qquad 22.1$$

where θ = volumetric water content (cm^3/cm^3); t = time (min); k = unsaturated hydraulic conductivity (cm/min); S = soil moisture tension (cm water), positive in unsaturated soil; z = depth (cm) and s = sink term (1/min). Eqn 22.1 is solved by an explicit forward difference method. The water content - tension relationship is treated by a modified form of the analytical function given by Brooks and Corey (1964) and possible hysteresis effects are ignored. The unsaturated hydraulic conductivities are calculated from the water retention curves and measured values of saturated hydraulic conductivity as proposed by Mualem (1976).

For the upper boundary condition submodels are available for precipitation, snow dynamics, interception and evapotranspiration. Measured precipitation is corrected for wind, wetting and evaporation losses. The fraction of frozen water in the precipitation is calculated from the temperature. The snow routine has separate accounts for liquid water and total water equivalent and the accounts are connected by a melting-freezing function. Evapotranspiration is calculated as a sum of actual transpiration and wet surface evaporation. The interception is treated as a threshold storage for the whole vegetation. The potential transpiration is calculated by the Penman-Monteith equation. The root water uptake is treated as a sink term and the potential transpiration may be reduced due to low water content. The reduction is performed separately for each depth where the normalised root density is above zero. No water uptake by roots in the groundwater zone is allowed in the model.

If the whole soil profile is unsaturated, the lower boundary condition consists of a unit gradient percolation. If a groundwater level is present above the bottom of the profile, a groundwater outflow (q_{gr}) is calculated from:

$$q_{gr} = q_1 \max(0,\ z_1 - z_{sat}) + q_2 \max(0,\ z_2 - z_{sat}) \qquad 22.2$$

where q_1, q_2 = maximum peak and baseflow respectively; z_{sat} = level where the soil water tension is nil and z_1, z_2 = levels where peak and baseflow respectively ceases. q_1, q_2, z_1, z_2 are obtained from calibrations.

The heat flow is calculated as the sum of conduction and convection using the general heat flow equation solved by an explicit forward finite difference method. Air temperature is given as the upper boundary condition when snow is absent. During situations with snow cover the thermal properties of the snow are taken into account. The lower boundary can be a temperature or a constant heat flow from below. The different thermal properties for organic and mineral soils, different water contents and unfrozen and frozen conditions are considered. Soil frost treatment is based on a function for freezing point depression and on analogy between freezing-thawing and drying-wetting capillary processes. Unfrozen water below zero degrees is thus associated with a soil moisture tension and an unsaturated conductivity when interaction between heat and water flow is allowed. Freezing gives potential gradients which force water to flow according to the prevailing conductivity.

Climatic data from SMHI's stations at Rörsbo and Växjö were used as driving variables. The precipitation data were corrected for wind, wetting and evaporation losses by 7% for rain and 21% for snow. The lower boundary condition was given as an existing groundwater table and a groundwater outflow (cf. eqn 22.2). The soil profile was divided into a number of compartments to account for numerical requirements and for observed vertical heterogeneity. In the root distribution adopted, the deepest roots reach 1.5 m below ground and 85% were located above 0.5 m.

The model was tested against observed groundwater level data. A dry and wet period were chosen for the tests (Oct. 1970-Feb. 1972 and Nov. 1979-Oct. 1980). The tests were performed separately for winter and summer periods. The winter periods were used to settle the groundwater flow parameters q_1, q_2, z_1 and z_2 of eqn 2 (15.0 mm/d, 0.4 mm/d, 0.4 m and 3 m). The fit for the summer periods was obtained by adjusting the surface resistance and its seasonal variation in the Penman-Monteith equation.

It appeared to be impossible to get good fit for the dry summer 1971. The simulated groundwater table was too shallow and the recovery in autumn was delayed. Changes in the assumed root distribution improved the fit for this year but caused disagreements for wetter years. Another factor of importance for predictions of the groundwater level was the upward capillary water transport. Earlier simulations (Johansson, 1986) showed that this transport was very sensitive to changes in water retention properties at low tensions. By assuming a higher air-entry value (30 cm) for the deepest layer, in contradiction to estimations from the laboratory determination of the pF-curve (Fig. 22.3), the best agreement was obtained independent of year.

Groundwater recharge was in the simulations defined as the accumulated flow across the compartment boundary immediately above the groundwater table. For a long period of time, when changes in storage may be neglected, the groundwater recharge will be equal to the accumulated groundwater outflow.

22.2.2 Soil moisture budget model

The most commonly used model of the unsaturated zone, for estimation of groundwater recharge, is a single reservoir model (see, e.g., Penman, 1949; Grindley, 1967 and 1969; Eriksson and Johansson, 1978; Howard and Lloyd, 1979 and Rushton and Ward, 1979). Conceptually the size of the reservoir is meant to represent the amount of water contained in the root zone between the field capacity and wilting point. At every time step, precipitation minus evapotranspiration is added to the reservoir. When the reservoir is full, the surplus will go to groundwater recharge. The evapotranspiration usually takes place at potential rate until a certain soil moisture deficit is reached. The actual evapotranspiration then decreases and reaches zero when the reservoir is empty.

Some earlier comparative studies suggested an underestimation of the groundwater recharge calculated with this model and an inability to represent the recharge dynamics as reflected in groundwater level fluctuations (Smith et al., 1970; Fox and Rushton, 1976 and Kitching et al., 1977). Rushton and Ward (1979) tested different ideas of by-pass flow to the groundwater zone when the reservoir was not full.

Results from the model are of very limited value without calibration and validation, since there is a substantial uncertainty in input data (precipitation and potential evapotranspiration).

Considering the soil water retention and flow properties in the study area, and the groundwater level fluctuating from immediately below ground surface almost down to the bedrock, it did not seem possible to determine a single site reservoir with the physical representation outlined above. However, it was considered to be interesting to test and compare this commonly used method against the simulations by the one dimensional soil water flow model and against observed groundwater level fluctuations.

The soil moisture budget model was applied in the Rörsbo area, where simulations with the one-dimensional soil water flow model had been performed earlier. The soil moisture budget model was calibrated and compared with the SOIL simulation giving the highest recharge for the period 1970-1984. The same climatic input data were used. (1969, which was the first year with rainfall data available from Rörsbo, was excluded since the results from the soil moisture budget modelling might be biased by the choice of the initial value of water content in the soil moisture reservoir.)

The structure of the soil moisture budget model programme was taken from Nilsson (1983). The relation between potential and actual evapotranspiration was given by:

$$AE = PE(S/SFC)^{c} \qquad 22.3$$

where AE = actual evapotranspiration; PE = potential evapotranspiration; S = actual soil moisture storage; SFC = maximum soil moisture storage and C = coefficient obtained from calibration. The snow routine was based on a degree-day approach. The concepts of no groundwater recharge before the soil moisture reservoir was full and allowing a fraction of rainfall or snowmelt for recharge also when a deficit existed, were both tested. The fraction forming recharge depended on the deficit (Fig. 22.4). The latter concept is often adopted in runoff modelling (see e.g. Bergström, 1976).

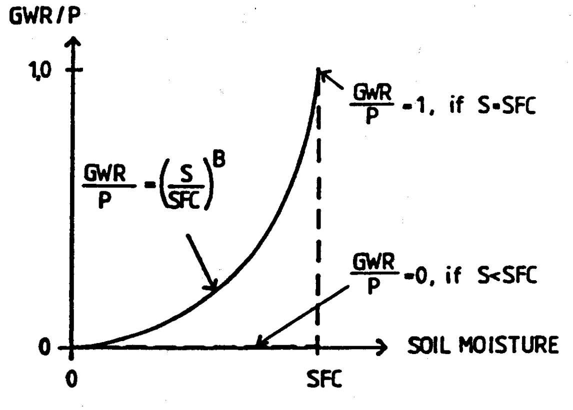


Fig. 22.4 Different concepts for the relation between groundwater recharge and water content in the soil moisture reservoir (GWR = groundwater recharge; P = precipitation and snowmelt; S = maximum water content in the soil moisture reservoir; SFC = maximum water content in the soil moisture reservoir; B = coefficient obtained from calibration)

22.3 Results

22.3.1 Estimation of groundwater recharge from the soil water modelling

The period 1969-1984 was simulated using the model parameters for groundwater flow and evapotranspiration obtained from the tests (simulation A). The sensitivity of the simulated groundwater level to displacement in the water balance, between evapotranspiration and groundwater outflow, was also tested for the same period. The parameter for maximum groundwater baseflow was put to 50 and 200% respectively of the value used in the first simulation (simulations B and C). These changes were compensated by simultaneous adjustments of the evapotranspiration according to the tests using the original calibration periods.

The correlations between simulated and observed groundwater levels were high for all three simulations (r = 0.92, 0.89 and 0.94) (Fig. 22.5). A closer examination revealed that during winter simulation C mostly gave a too rapid decline of the groundwater level, but a considerably better fit during summer 1979. For this same year, simulations A and B gave much too deep groundwater tables and the recoveries in autumn were too late. Generally all three simulations had a tendency to give too deep groundwater tables in summer and too late recoveries in autumn.

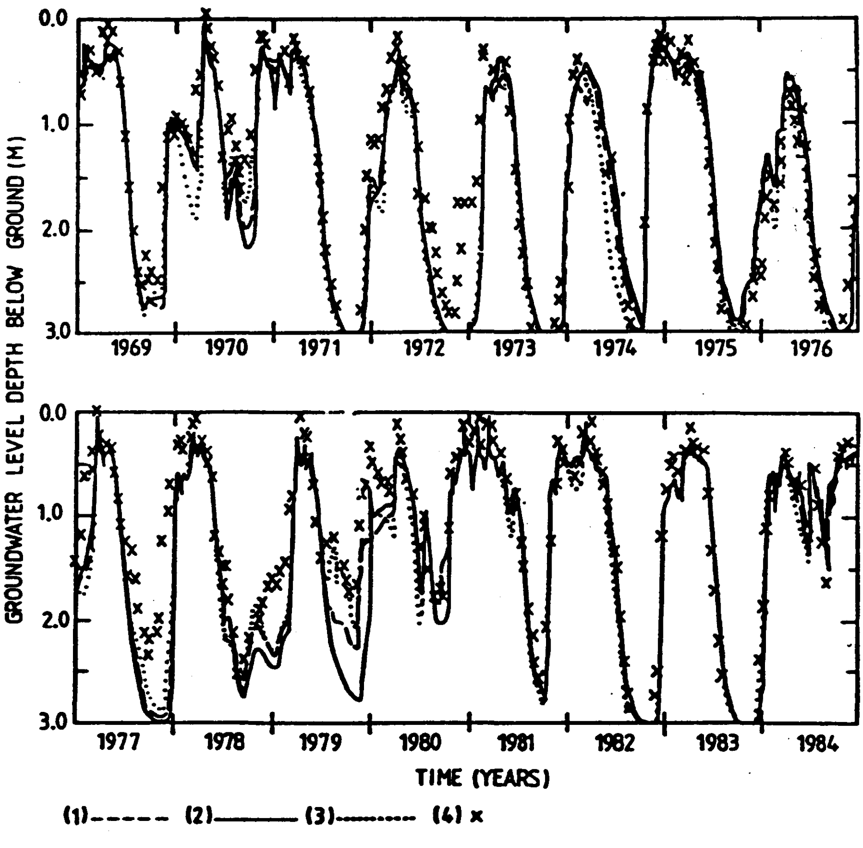


Fig. 22.5 Observed and simulated groundwater levels, 1969-84, at SGU's obs. tube No. 7, Emmaboda (1) sim. A, (2) sim. B, (3) sim. C, (4) observations

The different distributions between evapotranspiration and groundwater outflow in the simulations gave considerably different values for the groundwater recharge (Fig. 22.6). The average yearly evapotranspiration for the period was 440, 465 and 399 mm respectively and the resulting groundwater recharge 159, 134 and 197 mm.

All three simulations had March and April as the main recharge months, 37-53% of the yearly total. The simulations B and C and as a comparison, the runoff at SMHI's station in Kättilsmala (785 km²) in the river Lyckebyan are shown in Fig. 22.7. Simulation C gave the most evenly distributed recharge allowing a significant net recharge also during June to September. For June the other simulations gave a net transport from the groundwater due to capillary rise.

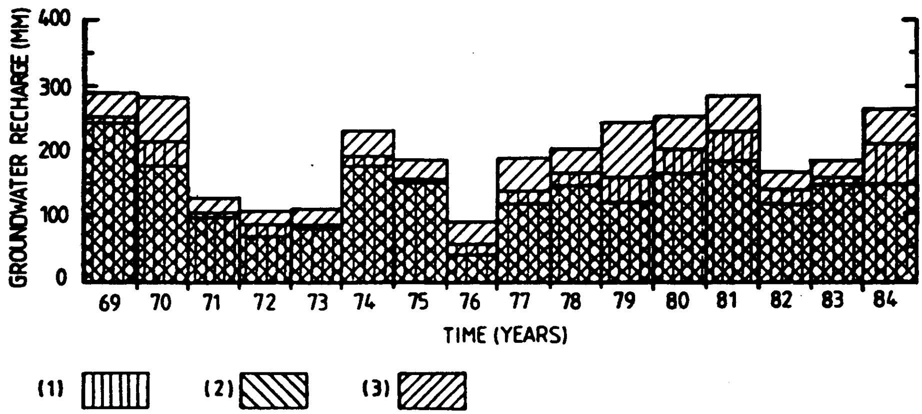


Fig. 22.6 Groundwater recharge, 1969-84, at SGU's obs. tube No. 7, Emmaboda
(1) sim. A, (2) sim. B, (3) sim. C

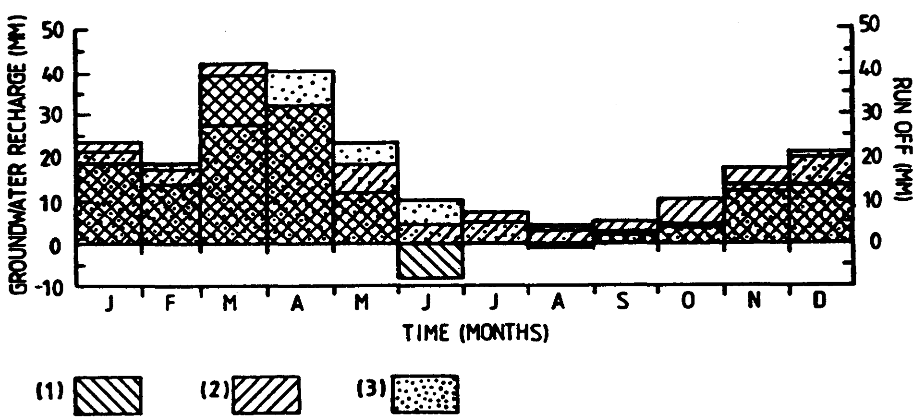


Fig. 22.7 Seasonal variation of groundwater recharge according to the simulations B and C and runoff at SMHI's station at Kätillsmala in the river Lyckebyan, (1) sim B, (2) sim. C, (3) runoff

The simulated upward transport from the groundwater zone was also appreciable, amounting to a yearly average of 29, 36 and 17 mm respectively for the simulations. During single years it was more than 30 mm also for the "least evapotranspiration simulation", and could for single months amount to 10 mm.

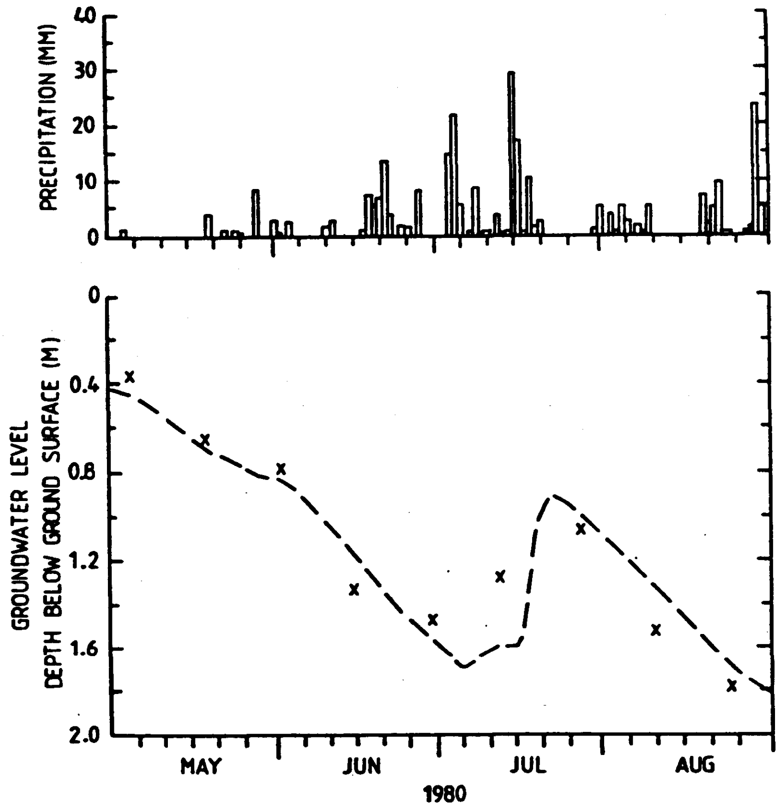
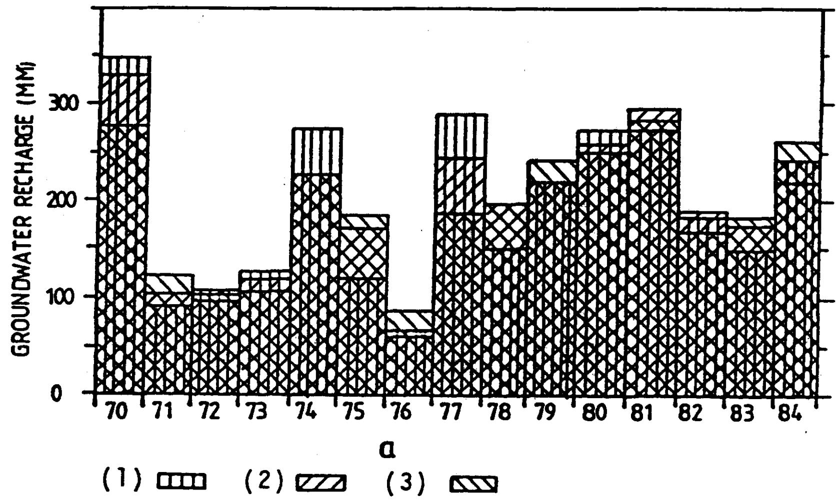


Fig. 22.8 Precipitation and observed (crosses) and simulated groundwater levels (dashed line), May-Aug. 1980 (SMHI's station at Rörsbo and SGU's obs. tube No. 7, Emmaboda)

From observations of rising groundwater levels, it was obvious that groundwater recharge also occurred during short periods most of the summers. Such a period occurred in July 1980 (Fig. 22.8). According to simulation A, the groundwater recharge was 15 mm during the period July 7 - 23. The groundwater outflow was 4 mm. The water table rise from 1.69 to 0.91 m below ground then was caused by a change in groundwater storage of 11 mm, corresponding to a specific yield of 0.014. The precipitation duriong July 1 - 23 was 118 mm. An autumn recovery may be exemplified by data from Oct. 24 - Nov. 8, 1974. A recharge minus outflow of 20 mm caused a water table rise from 2.89 to 0.78 m below ground resulting in a specific yield of 0.009. Studies of rises when the water table was more shallow e.g., for Dec. 8 - 12, 1980, gave 0.32 m of water table rise (0.47 to 0.15 m below ground) from 10 mm increase of groundwater storage, corresponding to a specific yield of 0.031.



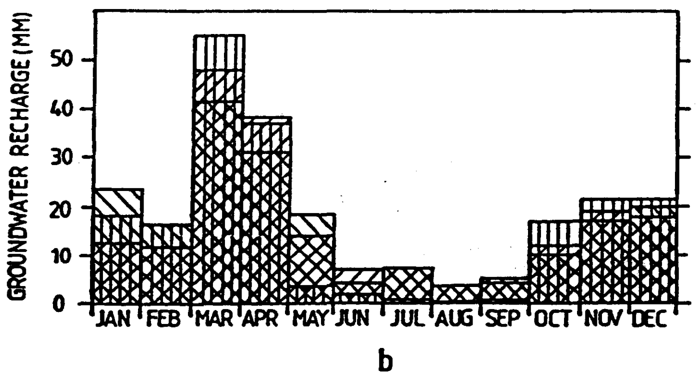


Fig. 22.9 Annual and monthly recharge, 1970-1984, at Rörsbo (SGU's obs, tube No. 7) simulated by two different variants of the soil moisture budget method and by one-dimensional soil water flow model (a) (1) soil moisture budget method with no recharge when reservoir deficits existed, (2) soil moisture budget method with a fractional recharge also when deficits existed, (3) one dimensional soil water flow simulation (SOIL)
(b) monthly averages of groundwater recharge for the period 1970-1984 (symbols as above)

22.3.2 Soil moisture budget model

The concepts with no groundwater recharge before the soil moisture reservoir was full and allowing a fraction of rainfall or snowmelt for recharge also when a deficit existed, were both tested. Several tests, with different sizes of the reservoir and the coefficient determining the fractional recharge, were performed. Results from one simulation based on each concept will be discussed, both giving approximately the same average annual recharge, for 1970-1984, as the SOIL simulation giving the highest recharge. Using the "no recharge before full reservoir" concept, the reservoir had to be about 150 mm, while 250 mm gave the best result together with a value of 5.0 for the coefficient B for the other concept. The average annual recharge was 194 and 195 mm respectivbly.

Studying annual and average monthly recharge, the simulation with recharge also when a reservoir deficit existed showed a better agreement with the SOIL simulation (Fig. 22.9a,b). Still substantial deviations were present for single years. The largest deviation, 25-30%, appeared for the years 1976-1977. No systematics in the deviations, coupled for example to dry and wet years, could be observed.

Looking closer to the dynamics, it was obvious that groundwater recharge during the summer was underestimated when no reservoir deficit was accepted. No recharge was simulated in situations when rising groundwater levels were observed. Recharge in autumn was also delayed compared to rising groundwater levels and sometimes completely missed before winter. Both summer recharge and autumn recovery were better reproduced in the simulations where recharge was allowed, also when the soil moisture reservoir was not full.

This case study is based on material published earlier by Johansson (1987, 1988).

23 RECHARGE IN THE MEHSANA ALLUVIAL AQUIFER, INDIA

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23.1 Introduction

This case study is chosen to illustrate how recharge estimates can be made for an alluvial aquifer. A variety of techniques are discussed including soil moisture balances, the use of tritium as a tracer and the water table fluctuation method. As the discussion proceeds it will become apparent that none of the techniques provides a complete understanding of the recharge processes; consequently a numerical model of the aquifer system is also developed to obtain an alternative estimate of recharge.

The Mehsana alluvial aquifer in Gujarat, western India, has been exploited for many centuries, Fig. 23.1a. Until the availability of power driven pumps, the quantities of water drawn from large diameter wells never caused a significant lowering of the water table. However, the availability of motor driven centrifugal pumps in the shallow wells resulted in the steady lowering of the water table, Fig. 23.1b. Due to the continuing fall in the water table, boreholes were drilled deeper into the aquifer; in many situations the piezometric heads (PH) of the deeper aquifers were found to be several metres above the water table, Fig. 23.1c.

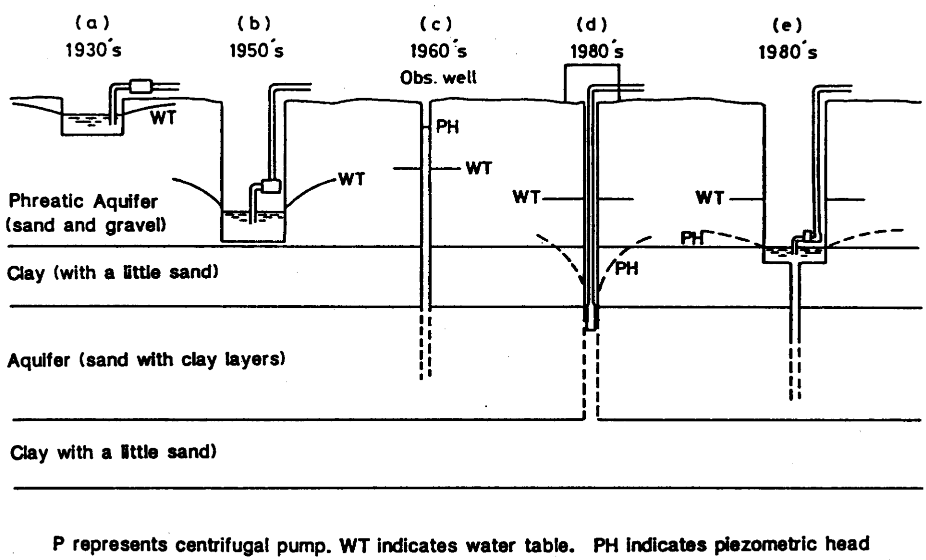

Fig. 23.1 Changing methods of exploiting the Mehsana aquifer, India

The success of this exploratory drilling and the small drawdowns observed during the initial pumping tests of the wells in the deeper aquifers suggested that the lower aquifers would provide a plentiful supply of water. Consequently, many of the dug wells were converted to dug-cum-bore wells and a large number of deep tubewells were also constructed. The optimistic predictions of the probable aquifer yield were not realised. As more wells were drilled the pumping levels fell; reductions of 50 m or more were not uncommon, Fig. 23.1d. Steady falls in the water table have also continued.

At one stage it did appear that the decline in water tables could be reversed when canal irrigation extended over part of the area. It was anticipated that the water lost from the irrigation system would travel quickly to the water table resulting in a rise in the water table. In practice, there has been little effect on the decline of the water table whilst water-logging is developing in certain areas.

This background has been presented to indicate that the over-exploitation in this aquifer is serious and that a realistic quantification of the recharge is essential before any steps can be taken to remedy the situation. Furthermore, the recharge to the aquifer cannot be determined by adopting only one technique; instead it is necessary to consider all available information in an attempt to gain an adequate understanding of the recharge processes.

23.2 Details of the aquifer system

The Mehsana alluvial aquifer extends over an area of about 3000 km² with the ground elevation falling from 180 m in the northeast to about 76 m in the southwest, Fig. 23.2 (Rushton and Tiwari, 1988). Average annual precipitation varies from about 900 mm on the higher ground to 500 mm over the lower ground. Maximum daily temperatures range between 42°C and 30°C with a mean annual potential evapotranspiration of more than 1500 mm. There are two main rivers, the Sabarmati which flows throughout the year due to reservoir releases and the non-perennial Rupen River. Canal irrigation is being developed over the northeastern part of the study area.

The aquifer system can be described as consisting of a phreatic aquifer and a complex sequence of sand, clayey sand and sandy clay lenses. A simplified cross section is contained in Fig. 23.3. From a number of observation wells which tap different depths within the aquifer it is possible to gain an insight into the current flow mechanisms. Fig. 23.4 shows that there are small fluctuations in the water table (Observation well P) with rises occurring during the rainy season of June to September and falls during the dry season. The response of the deeper aquifer piezometers, C1 and C2, is different. Recovery in levels commences in March and continues at a relatively constant rate until October; this response is primarily due to the variation in pumping for agriculture which is maximum during the period October to March but is far lower during the hot season and monsoon season from March to October.

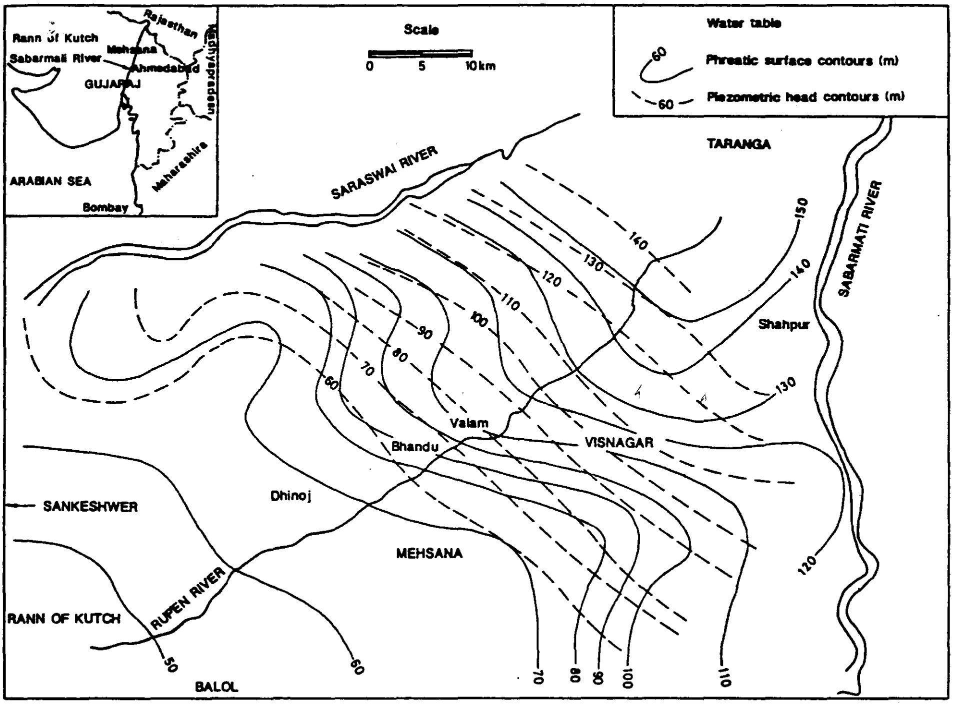

Fig. 23.2 Piezometric levels in the Mehsana aquifer

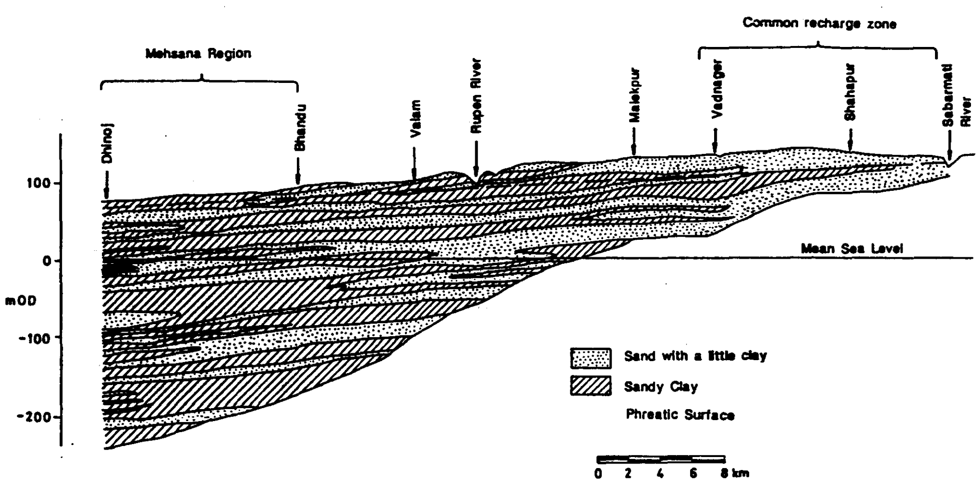

Fig. 23.3 Cross-section of the Mehsana aquifer

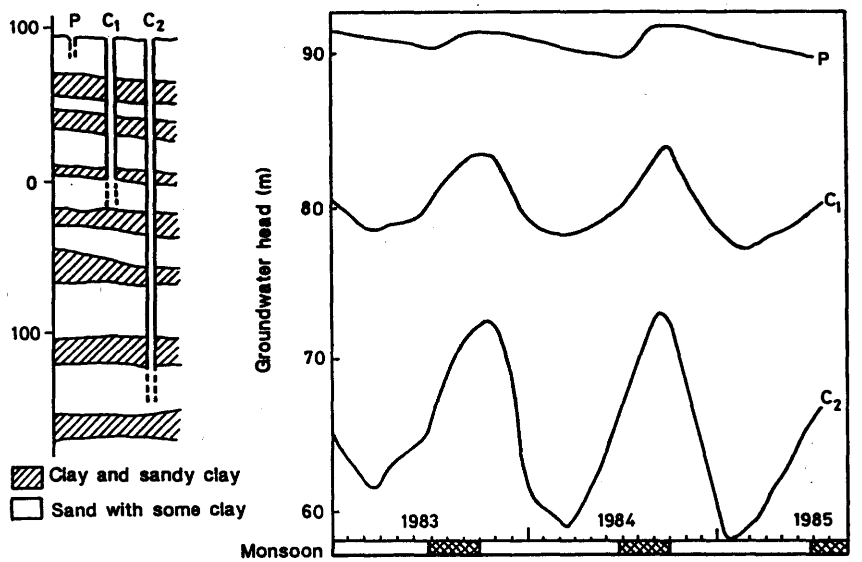

Fig. 23.4 Seasonal hydrographs at various depths

23.3 Sources of recharge

There are a number of potential sources of recharge to the aquifer. These include:

rainfall and irrigation recharge,
losses from canals and rivers,
losses from irrigated fields.

This section indicates how the potential recharge due to these various sources is estimated; the magnitude of the actual recharge which reaches the aquifer will be considered in a subsequent section.

23.3.1 Rainfall and irrigation recharge

As far as the ground is concerned there is little difference between rainfall and the application of water to the fields during irrigation. The convential method of estimating the recharge due to rainfall is to use a soil moisture balance technique as described in Section 11.4. Since there are only a few rainy days during the monsoon season and also because irrigation only occurs on a restricted number of days, it is essential to perform the calculation each day. The basis of the method is that a daily calculation is made of the conditions in the soil zone.

Conditions within the soil are expressed in terms of a soil moisture balance. When precipitation or irrigation occurs, some of the water may become runoff and the remainder of the water usually infiltrates into the soil zone. Of the water that enters the soil zone, some will be transpired by the crops and the remainder is stored within the soil zone. On days when there is no infiltration, the evaporative needs of the plants are met by the withdrawal of water from the soil zone. However if the soil zone is fully saturated and there is an excess of available infiltration, the soil may become free draining with the result that recharge to the aquifer equals the excess infiltration. For relatively coarse soils this is a reasonable approximation.

In the Mehsana region there are about 30 days each year when there is a high rainfall. For only about half of these is the soil fully saturated. Consequently, according to the soil moisture balance technique, recharge occurs on about 15 days each year. For the return flow to the aquifer due to irrigation, recharge only occurs when the irrigation is of such an intensity that the soil is fully saturated. This depends on the availability of the water; for farmers towards the top of the irrigation scheme this may occur during most waterings whereas the tail-end farmers rarely have enough water to fully saturate the fields.

23.3.2 Effect of farming practice

The above approach provides a reasonable estimate of the recharge for the coarser soils of the higher ground. However, in the lower ground of the Mehsana area, where the intensity of farming is far higher, the farming practice means that the above approach is not satisfactory. Many of the fields are used at times of plentiful water for flooded rice irrigation. Consequently, there is a less permeable layer within the soil which prevents the soil from becoming free draining. Another common practice is that during periods of heavy rain, the farmers ensure that any runoff is diverted to a low-lying field which is then used to grow flooded rice. This limits the recharge. Furthermore, because fields in the Mehsana area are irrigated using comparatively expensive well water, the farmers are careful in the use of water, pumping only sufficient water to prevent the soil from drying out but not enough to saturate the soil. Consequently, little of this irrigation water becomes recharge. Detailed recharge calculations show a wide variety of recharge intensities over the area. During drought years the rainfall or irrigation recharge over the entire region is effectively zero. During average rainfall years, the recharge over higher ground with the higher rainfall and coarser soils approaches 150 mm/y whereas in the lower lying ground the recharge is very small due to the farming practice. In years with a high rainfall, the recharge for the higher ground can exceed 300 mm/y and the recharge in the lower areas approaches 100 mm/y.

23.3.3 Recharge due to losses from canals

The conventional approach in estimating the recharge is to calculate the loss from a canal as a function of the wetted perimeter (Wachyan and Rushton, 1987). For main and secondary canals a typical value is 0.2 $m^3/d/m^2$, most of this loss becoming groundwater recharge. For a canal with a base of 2 m, depth of water 1 m and sides of 45° the loss equals 0.97 $m^3/d/m$.

This approach ignores two important features of the underlying strata which strongly influence the losses from canals, namely:

- the hydraulic conductivity including layered effects,
- the groundwater heads.

Losses from a canal do increase with the hydraulic conductivity and are strongly influenced by the nature of the layering and the groundwater heads. For a situation such as that illustrated in Fig. 23.5 where there is a low conductivity zone about 10 m below the bed of the canal, the loss to groundwater equals 1.91k $m^3/d/m$ where k is the hydraulic conductivity of the main aquifer. Where canals have been constructed in the more permeable areas, the hydraulic conductivity is at least 1.0 m/d, which is equivalent to a loss almost double the value quoted at the beginning of this section. Even if the canal is lined, the losses are likely to exceed 70 % of this value (Wachyan and Rushton, 1987).

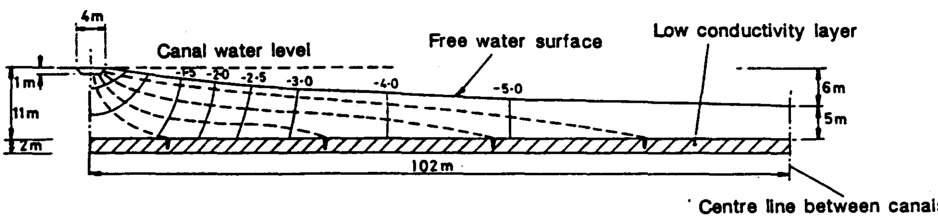

Fig. 23.5 Influence of hydraulic conductivity on canal losses

23.3.4 Losses from rice fields

Another example of losses to aquifers due to irrigation occurs with flooded rice fields. Detailed field and model studies (Walker and Rushton, 1984) have shown that substantial quantities of water can pass through the bunds of rice fields to the underlying aquifers. This occurs because the puddled layer does not continue under the bund.

For a typical rice field with an area of 2000 m^2 the loss can be expressed in the same form as an equivalent of water and equals approximately 40k m^3/d where k is the hydraulic conducivity of the bund. Therefore for soils with a moderately high hydraulic conductivity of 1.0 m/d the loss distributed over the area of the rice field is equivalent to 20 mm/d; this loss is significant both in terms of the poor efficiency of water use and in the high recharge which it produces. However, for clay soils the loss is small and indicates why the recharge due to irrigation in Central Mehsana is small. Furthermore, the loss from the rice field depends on the depth of water; the figures quoted above refer to a depth of water of 16 cm which is typical for canal irrigation schemes. When groundwater is used with daily waterings, the depth of water may only be 5 cm and the losses through the bunds then become significantly less.

23.3.5 Comments

This section has illustrated how physical models can be used to estimate the potential recharge through the soil zone or from canals or irrigated rice fields. To estimate the potential recharge over the whole area it is necessary to divide the area into zones and perform separate calculations for different rainfall intensities and different crop and soil types. Additional calculations are necessary to estimate the magnitude of recharge due to canals and flooded rice fields.

23.4 Estimating recharge by tracers

A variety of different tracer methods have been used to estimate the magnitude of the recharge to the Mehsana alluvial aquifer. Direct estimates of recharge have been made using environmental tritium and a greater understanding of inflows to and outflows from this aquifer have been made using ^{14}C, ^{32}Si and uranium.

23.4.1 Environmental tritium

Sukhija and Shah (1976) carried out an investigation into the change in environmental tritium profiles over an interval of two years. Three of their sites lie within or close to the study area, Fig. 23.2. They each relate to a different type of area:

- Balol is on the lower ground where there is intense agriculture, the soil is a sandy loam,

- Sankeswar has a sandy soil and is in the arid zone in the vicinity of the Rann of Kutch.

- Taranga is on the higher ground where the topsoil is a coarse sand,

Fig. 23.6 shows the tritium concentration in the precipitation in the south of the region; peak concentrations occurred in 1963-64 and have fallen steadily. Tritium profiles for the three sites are plotted in Fig. 23.7, values are quoted for 1967 and 1969. The figure also contains profiles of the soil moisture content. Two alternative methods of estimating the recharge are used. Each is based on the assumption that the tritium moves at the same rate as the water. Further, a piston flow mechanism is assumed in which the younger water causes the older water underneath to move downwards.

In the peak method, the movement of the peak due to the 1963/1964 rainfall is traced. The recharge is computed by finding the amount of water present from the surface downwards to the tritium peak position. This is compared with the precipitation since 1963 to the time of the investigation; the ratio of these two determines the recharge rate.

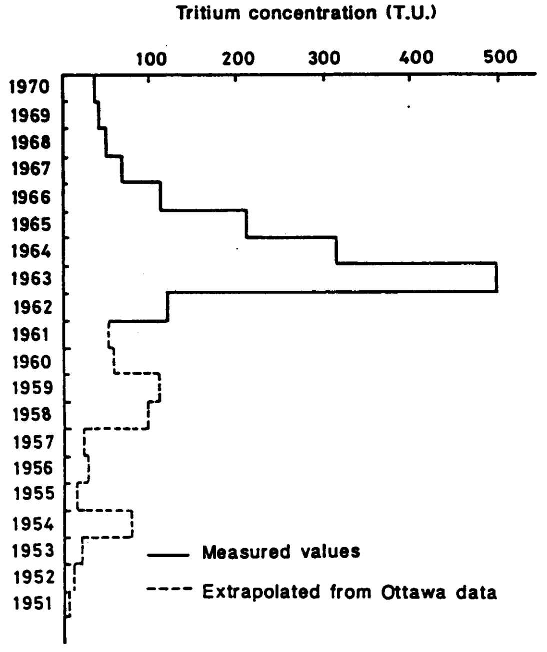

Fig. 23.6 Tritium concentrations in precipitation, Gujarat

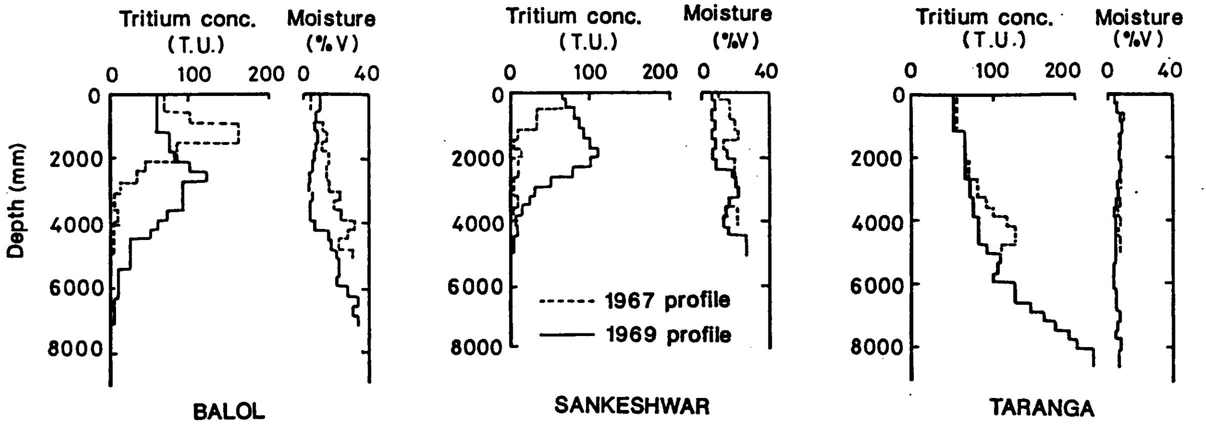

Fig. 23.7 Tritium profiles near Mehsana study area

In a second method, called the total tritium method, a balance between tritium fallout and tritium accumulated in groundwater is obtained; the difference is attributed to evapotranspiration and runoff. The ratio of tritium in the groundwater to total tritium in the precipitation equals the ratio between recharge and total precipitation. Tritium concentrations are corrected for radioactive decay and the calculation covers the period from 1952 to the date of the investigation. Recharge estimates for the period 1967 to 1969 are listed in Table 23.1.

This table shows that the two methods of estimating the recharge from the tritium profiles give similar results at Balol and Sankeshwar. The precipitation at Sankeshwar is less than at Balol and the recharge values show a similar trend. When expressed as a percentage of the rainfall, the recharge values are between 3-4%. This percentage is lower than the value that is normally used in India, but from knowledge of the field conditions such a low value is not unexpected.

Table 23.1 Recharge estimates for 1967 to 1969 using environmental tritium measurements

	Annual recharge (mm)		
Site	Peak method	Total method	Mean
Balol	28	25	26
Sankeshwar	17	13	15
Taranga	58	>27	-

An examination of the tritium profiles at Taranga indicate that the downwards movement is so rapid that it is difficult to use the total tritium method. Even the use of the peak method is open to question since the profile hardly extends to the peak. The recharge estimate of 58 mm/y is higher than at the other sites but it is still smaller than estimates indicated by other methods for the higher ground with the coarse soil.

As an alternative to the environmental tritium method described above, it is possible to inject tritium in the ground. Injected tritium studies which have been carried out in the study area are described by Gupta and Sharma (1984). Recharge rates were estimated to be 9% of the rainfall.

Tritium measurements provide point estimates of the recharge. Usually they refer to areas without growing crops and furthermore, a site is usually selected where the soil profile is relatively uniform. Consequently the results are unlikely to be representative of most of the study area which is under agricultural production. Nevertheless, the tritium technique does provide useful information for overall assessment of the recharge.

23.4.2 Other tracers

Tracers have proved to be useful in learning more about the likely flow paths within the aquifers; the results can provide information to check the recharge estimates. Gupta et al. (1981) use the ratio of ^{32}Si and ^{14}C which have half lives of 300 and 5730 years respectively. Different ratios of these parameters can be used to estimate whether the water deeper in the confined zones originates mainly from horizontal flow from more distant recharge zones or from vertical flow through the less permeable layers. Vertical flow is found to be very important in the Mehsana alluvial aquifer.

Uranium isotopes have been used to learn more about the interaction between phreatic aquifers and the Sabamarti River. Information has been gained about the flows from the aquifer to the river; this can provide insights about the possibility of the river recharging the aquifer. Borole et al. (1979) used ^{238}U concentrations and the $^{234}U/^{238}U$ activity ratio in boreholes and in the river to estimate the flows from the aquifer to the river. Using a simple mixing model, estimates have been made of the flow from the aquifer to the Sabarmati River. Along one stretch of the river the measured uranium concentrations in the river were found to be less than those for the river water flowing into that stretch. This suggests that water flows from the river to the aquifer but it was not possible to estimate the magnitude of the flow.

23.5 Water table fluctuations

Estimates of the recharge are frequently made on the rise in the water table during the rainy season. The concept is simple; the water table rise is multiplied by the specific yield to give the recharge. Fig. 23.8 shows the water table fluctuation in a dug well in the Mehsana region; values are quoted for May which is before the start of the monsoons, October which coincides with the end of the monsoons and January which is the middle of the dry season. Water table rises for years with higher rainfall are generally about 2 m and during years with lower rainfall the rise is closer to 1 m.

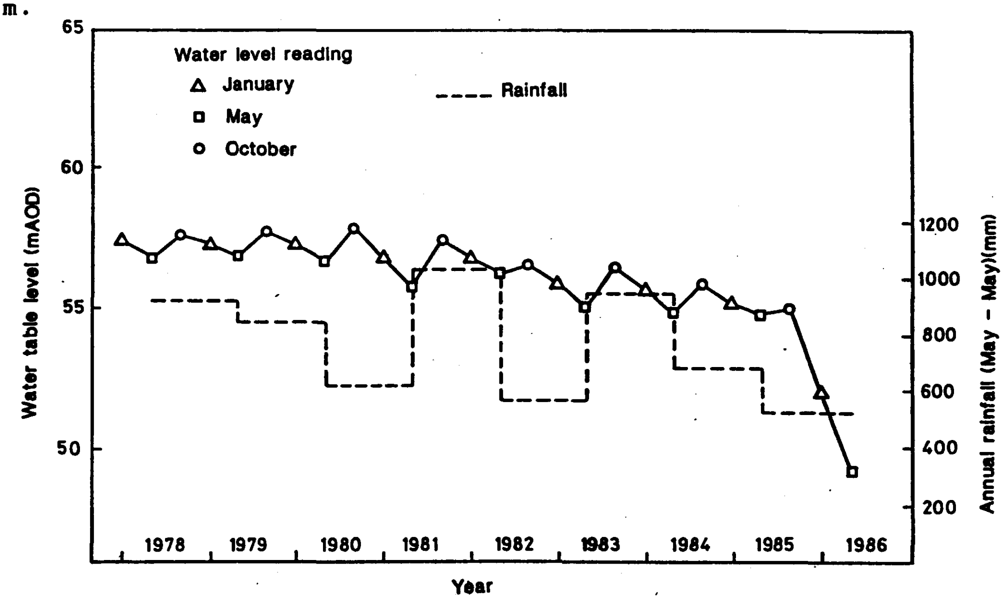

Fig. 23.8 Groundwater level fluctuations in a dug well

The next step is to estimate the specific yield. Pumping tests have generally led to the value of 0.02 to 0.05 whereas text books (Todd, 1980) suggest drainable porosities of 0.15 to 0.2 for sandy or sandy clay soils. However, when note is taken of the unsaturated conditions within the zone of water table fluctuation the actual meaning of the specific yield becomes less clear (Section 16.2). If water is pumped from shallow wells during the monsoon season, this must be added to the recharge estimate.

Despite the difficulty in determining both the magnitude of the water table fluctuation and the appropriate value of the specific yield, the water table fluctuation method, together with estimates of other sources of recharge is often used as a means for determining the recharge to an aquifer. Table 23.2 refers to a recent attempt at assessing the recharge components for the Mehsana alluvial aquifer. Two areas are considered, Visnagar which is on the higher ground with a more sandy soil and Central Mehsana where soils have a higher clay content.

Table 23.2 Recharge estimation for two areas in the Mehsana alluvial aquifer. Results are quoted as 10^6 m³ per year for a typical year

Recharge source	Visnagar	Central Mehsana
1. Monsoon recharge	89	45
2. Discharge from shallow wells during monsoon	24	18
3. Canal losses	6	0
4. Return flow from irrigation	7	0
5. Seepage from lakes	4	5
6. Influent seepage from rivers	15	1

These quantities are estimated in the following manner:

(1) Monsoon recharge for the Visnagar area is calculated from the area of 488 km^2, an annual water table fluctuation between 1979 and 1983 of 5.55 m and a specific yield of 0.033. For Central Mehsana the corresponding figures are 398 km^2, 3.78 m and 0.03.

(2) Discharge from the shallow wells can be estimated from the number of wells and the method used for withdrawing the water.

(3) Canal irrigation has only extended to the Visnagar area. Branch canals have a total length of 25 km with an average wetted perimeter of 10 m whilst the distributaries extend over 32 km with a wetted perimeter of 6 m. For an operating season of 250 days and an assumed loss of 0.056 m/d per unit area of canal base or sides, the total loss is calculated as 6.2×10^6 m^3 per year.

(4) Return flow from irrigation is estimated as a proportion of the canal water delivered to the farmers' fields.

(5) Seepage from lakes could be a major source of inflow. However, because there is an accumulation of silt in the beds of lakes, the loss is assumed to be only 0.005 m/d per unit plan area for a period of 60 days. For Visnagar the area of the lakes is 13 km^2 and for Central Mehsana the area is 17 km^2.

(6) Losses from rivers are estimated using Darcy's Law. The Rupen River passes through both the areas. In the Visnagar region the length of the river is 12 km and assuming a transmissivity of 680 m^2/d and a hydraulic gradient of 0.005, the total loss during 365 days is 15 x 10^6 m^3 per year. For the Central Mehsana area the corresponding values are 8 km, 275 m^2/d and 0.006.

Many comments could be made about these estimates. For example, the values of the water table fluctuations are difficult to obtain when results such as those of Fig. 23.8 are examined. In fact, the quoted values of 5.55 and 3.78 m are based mainly on the extreme fluctuation in one year. The quoted values for the specific yield are also open to question since they are determined from relatively short pumping tests rather than on the longer term drainage due to the fall in the water table.

Canal losses are almost certainly an underestimate for years when there is sufficient water to maintain flow for as long as 250 days. In addition return flow from groundwater irrigation is significant in the more permeable Visnagar area. Seepage from lakes as quoted in Table 23.2 also appears to be underestimated; most of the lakes empty in two to three months which can only be accounted for by a higher loss to the aquifer.

Of all the calculations, the losses from the rivers is probably the least reliable. The calculation assumes that water flows horizontally from a river which fully penetrates the aquifer. In fact the river is perched above the aquifer with the true water table at 10 or more metres below the river. The actual flow pattern is similar to that of Fig. 23.9. Furthermore, the actual flows in the river are small or zero during much of the dry season and therefore the values quoted in Table 23.2 are up to ten times the actual values.

Further critical comments could be made about the results of Table 23.2. However, the important positive comment is that Table 23.2 does provide an attempt to identify and quantify all the possible sources of recharge. The main inadequacy is that some of the calculations have been performed without identifying clearly the physical nature of the flow processes.

There is, however, a further limitation in the approach discussed above for estimating the rainfall recharge. This approach assumes that during the recharge season the aquifer acts as a "bucket" which does not leak. In practice there are other inflows and outflows. These include:

- lateral flow in the aquifer,
- abstraction from the unconfined aquifer (this has been included in Table 23.2),
- vertical leakage from the water table aquifer through clay layers into the underlying aquifers. This vertical flow is indicated by the difference between the groundwater heads in the shallow and deep aquifers, Fig. 23.4; this vertical leakage occurs throughout the year.

From the above discussion it is clear that the use of the rise in water table during the rainy season is likely to lead to unreliable predictions of the rainfall recharge. However, the long-term trend in the water table elevation, which for the well hydrograph of Fig. 23.8 is about 0.3 m/y, can provide valuable information about the aquifer response. This will be considered with the use of mathematical models.

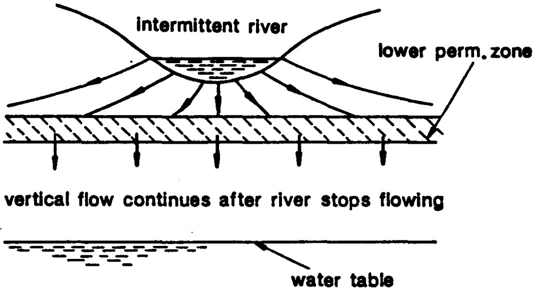

Fig. 23.9 Recharge from a perched river

23.6 Actual and potential recharge

The difference between actual and potential recharge becomes clear when note is taken of the change in aquifer conditions as canal irrigation extended to part of the area during the last ten years; the development is not yet completed. Since the losses from canals to the aquifer are high, it was anticipated that a rapid recovery would occur in the water table levels. In the areas where the aquifer is a coarse sand, some recovery has been noted. However, elsewhere the water table continues to decline at the same time as waterlogging is occurring in the vicinity of the canals.

The reason for this behaviour is that clay layers in the unsaturated zone restrict the downwards movement of water with the result that water is ponding on these clay layers (Rushton, 1986). Therefore actual recharge reaching the water table does not equal potential recharge which arises from losses from the canal irrigation.

Fig. 23.10 illustrates a typical situation. The potential recharge due to the canal irrigation is at least 5 mm/d. Between the ground surface and the water table there is a sandy-clay layer which is 2 m thick of hydraulic conductivity 0.0008 m/d. Water collects above the clay layer to cause a perched water table 1.5 m above the clay layer. The vertical hydraulic gradient is therefore 3.5/2.0 and the actual recharge can be calculated from Darcy's law to be 1.4 mm/d.

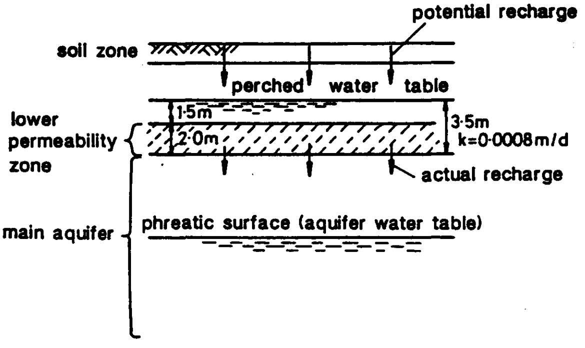

Fig. 23.10 Potential and actual recharge in an alluvial aquifer

This difference between actual and potential recharge is highlighted by the conditions in the vicinity of canals but it does occur elsewhere in the study area. Therefore all the previous estimates, whether obtained from water balance calculations or from tracers, must be modified if there are less permeable layers present in the unsaturated area.

23.7 Recharge estimated from mathematical models

The previous sections have described attempts at estimating the recharge using a variety of techniques, and with each there is some uncertainty. This uncertainty is increased by the influence of the clay lenses which limit the actual recharge and also create significant vertical gradients within the aquifer. This section presents a brief description of a mathematical model representing flow conditions in a typical vertical section which can be used to provide an indirect estimate of the recharge.

The mathematical model represents a section of the Mehsana aquifer, 10 km wide and extending for a distance of about 60 km in the general direction of the flowlines; the outline of the modelled area is shown in Fig. 23.2. From all the available borehole data a representative cross section is devised, Fig. 23.3. This cross-section is modelled using a finite difference technique; the mesh divisions are chosen so there are roughly equal numbers of mesh intervals in the horizontal and vertical directions. Hence the horizontal mesh spacing is 2 km and the vertical mesh spacing is 15 m. The coefficients of the finite difference equations depend on the relative proportions of sand and clay in each interval.

The time-instant technique is used whereby the model represents the average conditions within the aquifer during a particular month. For instance, during October the abstraction has not built up to the peak and it is therefore approximately equal to the average annual abstraction rate. In this model, this abstraction is withdrawn from more permeable layers with three withdrawal points on each vertical line of nodes. The boundary conditions are chosen as follows:

- no flow crossing the lower boundary

- saline water is present in the south-west; there is no pumping and therefore this boundary approximates to a no-flow condition

- during the month the movement of the phreatic surface (or water table) is small compared to the groundwater head variation within the confined aquifer. Therefore the phreatic surface is represented as a specified head.

This condition on the phreatic surface required further consideration. The term phreatic surface is used instead of water table to emphasise that this is the surface below which the whole of the aquifer is saturated. In an earlier model this boundary was represented as having a known flow. The consequence was that the groundwater heads were too high leading to a phreatic surface which was above ground level. This was one of the first indicators that the actual recharge at the phreatic surface was less than most of the estimates of the potential recharge. Therefore the alternative approach of representing the phreatic surface as a specific head is adopted with the quantity of water drawn from the phreatic surface calculated from the resultant vertical head gradient.

When these conditions are imposed, the numerical model reproduces the vertical gradients of the groundwater heads. For example, at a piezometer nest with the phreatic surface at 83 mOD, the measured groundwater head in a piezometer 220 m below ground level is 49.2 m and the value according to the model is 52.6 m.

It is also possible to determine from the model the vertical flows <u>from</u> the phreatic surface; Fig. 23.11 represents the vertical flows in the Mehsana area. The average value of this vertical flow is 0.65 mm/d. This is the amount of water moving downwards from the phreatic surface. It originates from:

- recharge to the phreatic surface,
- water released due to a fall in the phreatic surface.

The second of these components can be deduced from the long-term average fall in the phreatic surface, which in the Central Mehsana area is approximately 0.9 m each year. With a long-term specific yield of 0.2, this annual fall is equivalent to 0.50 mm/d. Consequently the recharge actually reaching the phreatic surface is the difference between the vertical flow of 0.65 mm/d and the water released from storage at the phreatic surface of 0.50 mm/d. This estimated recharge of 0.15 mm/d or 55 mm/y is of the same order as the estimate deduced from the tracers. As with all the methods described previously, there are major uncertainties about this approach. In particular, the recharge estimate depends on an assumption about the amount of water released from storage. This uncertainty could be overcome by developing a time-variant numerical model, but for the complex Mehsana alluvial aquifer such a model would involve further uncertainties.

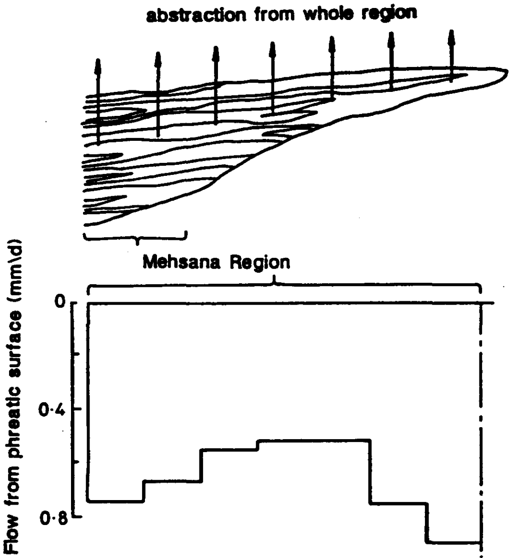

Fig. 23.11 Vertical groundwater flows, Mehsana aquifer

23.8 Discussion

This case study is concerned with the recharge estimation for a complex alluvial aquifer. Due to the intensive irrigation over much of the area, recharge can occur due to both precipitation and return flow from irrigation. Some recharge can also occur from rivers.

A wide variety of methods have been used. Each provides valuable information but no one method provides a complete answer. The one certain conclusion about the Mehsana alluvial aquifer which is confirmed by each of the methods is that the quantity of water pumped from the aquifer is far higher than the recharge.

24 ANALYSIS OF LONG DURATION PIEZOMETRIC RECORDS FROM BURKINA FASO TO DETERMINE AQUIFER RECHARGE

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24.1 Introduction

The marked interannual variability of rainfall in the African Sahel causes still greater variability in aquifer recharge and piezometric levels. The severe drought encountered since the 1970's has often resulted in substantial drops in water level, despite slight seasonal recovery. Long-duration observation data are only rarely available although data sequences covering 5 to 10 years can be used to analyse the possible evolution in water level for various rainfall scenarios.

An observation well was sunk to a depth of 20 m in a granite aquifer in Ouagadougou (Burkina Faso). The well is screened from 6 to 20 m, and taps 5 m of granitic sand, 4 m of weathered granite, and 5 m of fresh granite. It has been monitored since 1978 by the ICHS, which has related data covering an eight year period from 1978 to 1985. These data were subjected to detailed analysis using a lumped-parameter model, the aim being to extend the data sequence.

24.2 Available data

24.2.1 Piezometric levels

Water levels in the ICHS observation well in Ouagadougou were manually monitored by the ICHS between 1978 and 1985 (Diluca and Muller, 1985). Fig. 24.1 shows a continuous fall of 0.45 m/y in the average levels. A rise in level of about 0.80 m every year is nonetheless recorded for several months in autumn before levels fall again.

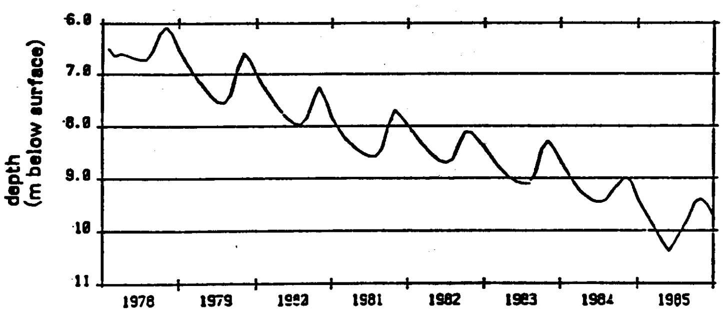

Fig. 24.1 Evolution in water level in the ICHS observation well in Ouagadougou (Burkina Faso) from 1978 to 1985

24.2.2 Hydrodynamic characteristics

No data are available, only one very short duration pumping test having been undertaken in May 1978 near the study area. In particular, the storage coefficient (or effective porosity) is not known, although it can be estimated at between 1 and 8% on the basis of the rock type.

24.2.3 Rainfall

Daily rainfall figures from 1959 to 1985 (27 years) were provided by ORSTOM and ICHS. Average annual rainfall for the 1959-1985 period was 825 mm, but only 690 mm from 1978 to 1985. The number of days of rainfall was 73 per annum for the 1959-1985 period, but only 60 per annum from 1978 to 1985. Monthly rainfall figures are also available for the 1929-1958 period.

24.2.4 Potential evapotranspiration (PET)

The monthly values applied for the model were calculated in Ouagadougou using Turc's formula. Only the interannual means (for an unspecified period) could be taken from an atlas (Lemoine and Prat, 1972). The total annual PET is 2084 mm.

24.3 The lumped parameter model

BRGM's model was used to calculate the balance of rainfall, potential evapotranspiration, runoff and infiltration. This lumped parameter hydrological model makes it possible to produce a local balance in daily time-steps and to calculate the actual evapotranspiration (ETR), runoff, infiltration and spot water-table level. The model is described in detail by Roche and Thiery (1984), the basic principles being briefly reviewed below.

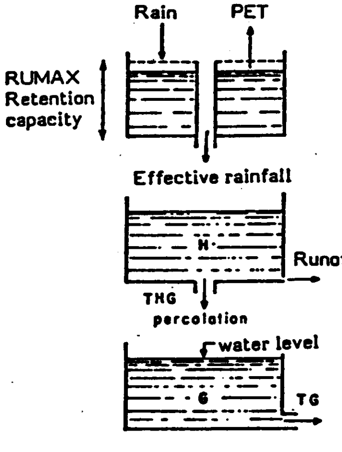

Fig. 24.2 Principle of the GARDENIA model used to simulate piezometric levels

The GARDENIA model consists of three superimposed layers (Fig. 24.2). The first (RU) is characterised by its retention capacity (RUMAX, or maximum soil moisture deficit), and represents the retention effect in the first few metres below the soil surface. This layer is supplied by rainfall, and is emptied by evapotranspiration. No runoff or infiltration occurs before this layer is saturated. This layer takes account of the effects caused by the interception in surface depressions, and schematizes the "valve effect" of unsaturated soil depending on the degree of humidity.

The second layer (H) is characterized by two parameters, a half-percolation time (THG), and an equal runoff-perçolation level (RUIPER). This layer transfers water to the water table through the unsaturated zone, and controls distribution between runoff and infiltration: the higher the level in this layer (as a result of heavy rainfall), the greater the runoff proportion. When the level of the layer is the same as the equal runoff-percolation level, infiltration equals runoff; when the level of the layer is lower infiltration is greater than runoff.

The third layer (G) is characterised only by the half recession time, and represents an exponential aquifer recession. Thiery (1985) has shown that this scheme corresponds for practical purposes to an aquifer bounded on one side by an impermeable rectilinear barrier and on the other by an imposed-level rectilinear barrier. If the observation well is positioned sufficiently far away from the imposed-level boundary, the piezometric level is deduced from the level G in layer G by the formula:

$$PL = G/STO + BL$$

where PL is the piezometric level, G is the level in the layer, STO is the unconfined storage coefficient or specific yield or equivalent effective porosity, and BL is the base level.

Four parameters are therefore to be determined:

(1) the retention capacity (RUMAX), which alone controls the value of actual evapotranspiration derived from potential evapotranspiration (PET)

(2) the half-percolation time (THG), and the equal runoff-percolation level (RUIPER) which control the runoff proportion and the delay-time between excess water in the soil and a rise in the aquifer level

(3) the half-recession time (TG) which governs the rate of aquifer recession

and also two amplitude parameters:

(1) the base level (BL)

(2) the storage coefficient or effective porosity (STO).

In order to calculate the balance under valid conditions, sequential data (not necessarily simultaneous) for the following parameters should ideally be available:

(1) runoff

(2) piezometric level and storage coefficient for the unconfined piezometric fluctuation surface.

The parameters of the model are in this case adjusted in order best to reproduce the two sequences of data. In practice, as is the case in the example cited:

(1) data for the runoff are not always available (except where runoff can be assumed to be nil or negligible)

(2) the unconfined storage coefficient is inadequately known (or sometimes a confined aquifer storage coefficient, quite unrelated to aquifer recharge).

It may be thought, a priori, that data on the piezometric level may be used alone to determine the retention capacity (RUMAX). If the value assumed for the model is too low, the levels calculated will affect the rainfall sequences too soon and too frequently. If this value is too high, the effect of piezometric levels will occur too late and in some cases not at all.

This distribution of runoff and infiltration for effective rainfall can therefore be calculated in unique solution form using the non-linear scheme for reservoir H, which does not give a fixed infiltration percentage but smooths the effect of heavy effective rainfall. Regulation of this smoothing of heavy effective rainfall (by means of the RUIPER parameter) will therefore regulate distribution between infiltration and runoff.

Infiltration will thus be transformed into variation in the piezometric level by the half-recession time (TG), the amplitude being inversely proportional to the storage coefficient.

This type of calibration poses no problems where piezometric levels vary rapidly in relation to sequences of isolated and non-periodic rainfall. In practice, as is discussed in Section 24.4, piezometric levels are often cushioned from the direct effects of rainfall and react only to a true "rainy season" rather than to a sequence of isolated rains. The sequence of water levels therefore describes a pseudo-sine curve in response to the pseudo-sine curve for the smoothed rainfall. Calibration consists of reproducing the pseudo-sine curve for water levels, with a time-lag in relation to the rainfall curve. If the storage coefficient is inadequately known, the variations in level cannot easily be reduced to millimetres of recharge and, as will be seen, a unique solution for calibration is likely to be no longer possible.

24.4 Calibration of model

It has been seen that evapotranspiration is only governed in the model by a single parameter: the soil retention capacity (RUMAX). This parameter is therefore essential because it allows determination of the effective rainfall, or rainfall minus ETR. This effective rainfall is itself distributed between runoff and recharge by the RUIPER parameter (associated with the half-percolation time THG). The rate of recession is only governed by the half-recession time TG.

It thus emerges that two parameters are of fundamental importance for the calculation of recharge:

(1) The soil retention capacity (RUMAX)

(2) The equivalent-runoff level (RUIPER).

24.4.1 Determination of soil retention capacity

Six calculations were made by fixing the soil retention capacity at 0, 10, 20, 40, 70 and 100 mm respectively. Fig. 24.4 shows the value for effective rainfall (runoff + infiltration) obtained for the 1978-1985 period as a function of retention capacity. Fig. 24.3 shows that the best results are obtained for a retention capacity of less than 40 mm. At capacities of 40, 70 and particularly 100 mm, the quality of calibration deteriorates, some years (1984 and 1985) displaying no rise in level. Fig. 24.4 shows that the effective rainfall calculated is highly dependent on the retention capacity, decreasing from 482 mm/y for nil retention to 196 mm/y for a retention of 20 mm.

24.4.2 Determination of runoff proportion

The runoff proportion is governed by the RUIPER (equal runoff-percolation level) and THG (half-percolation time) parameters. Six calculations were made for three retention-capacity values (10, 20 and 30 mm) associated with two storage coefficient values (1% and 4%). Fig. 24.5 shows that the six simulations are acceptable, all typified by correlation coefficients ranging from 0.965 to 0.980. Ajustments obtained with a storage coefficient of 1% or of 4% are equally satisfactory, the optimum retention capacity being 10-30 mm in both cases. The overall storage coefficient cannot therefore be determined more precisely by simple analysis of natural variations in water level.

Fig. 24.6 shows variation in surface runoff associated with the six hypotheses used in calculation. It is very clear that runoff varies markedly from hypothesis to hypothesis, being much more sporadic for a retention capacity of 30 mm than for a retention capacity of 10 mm, and being about 50% higher for a storage coefficient of 1% than for a storage coefficient of 4%.

Assuming a storage coefficient of 1%, the average annual recharge is 23 to 45 mm/y when retention capacity is chosen between 10 and 30 mm. Assuming a storage coefficient of 4%, the computed annual recharge would be 75 to 142 mm/y.

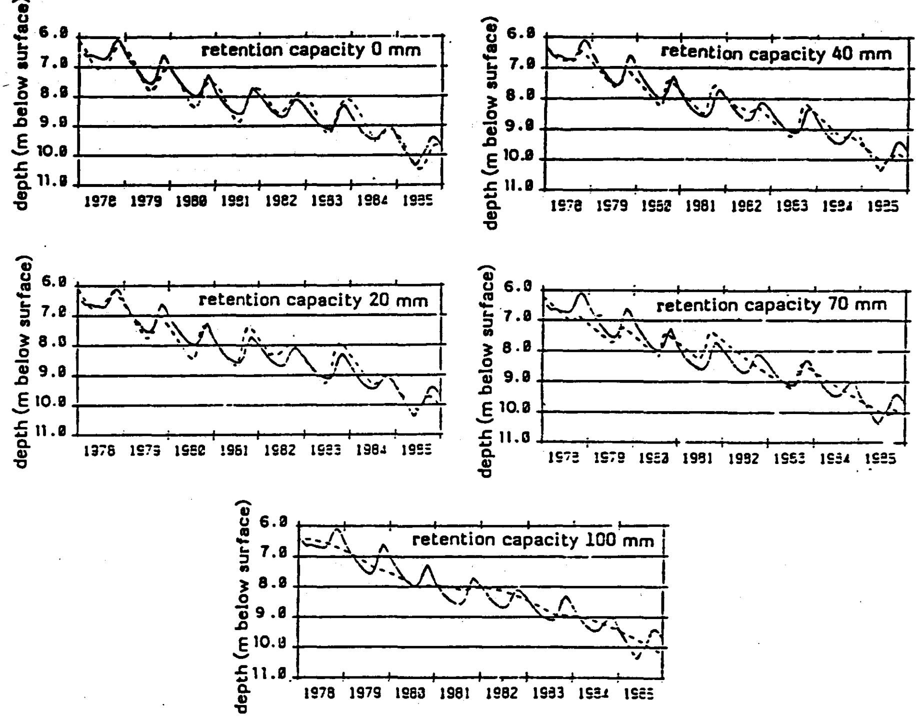

Fig. 24.3 Simulation of water levels for a retention capacity ranging from 0 to 100 mm

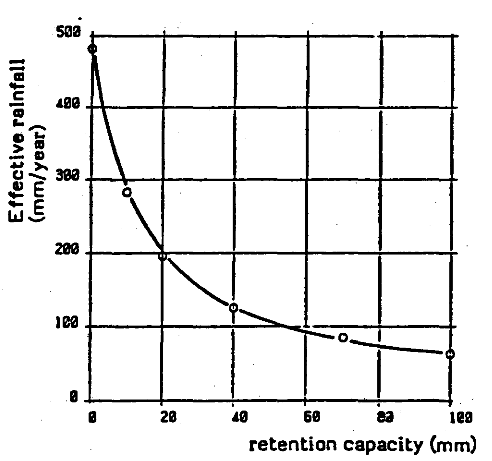

Fig. 24.4 Relationship between the effective rainfall and the retention capacity assumed for the model

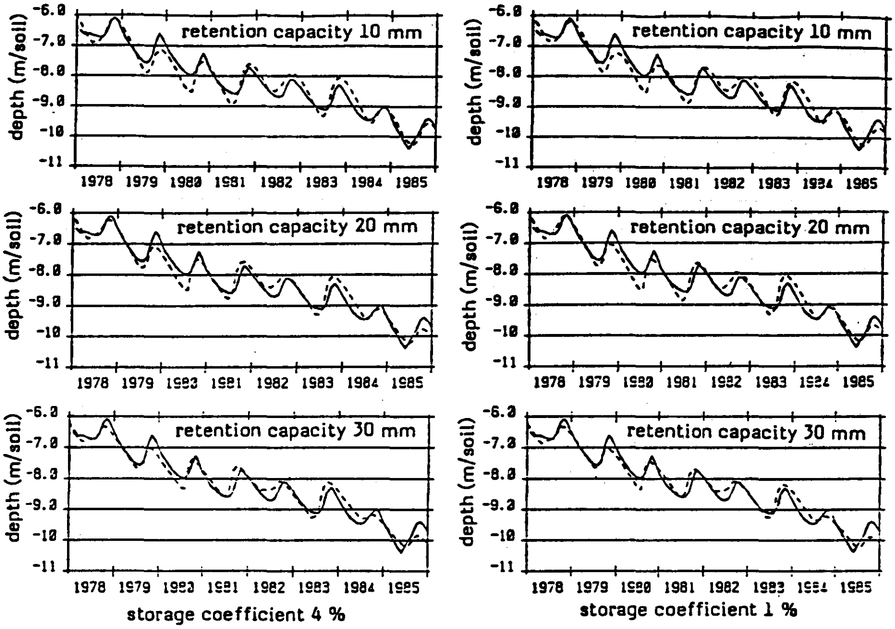

Fig. 24.5 Simulation of piezometric level for three retention-capacity values and two storage coefficients

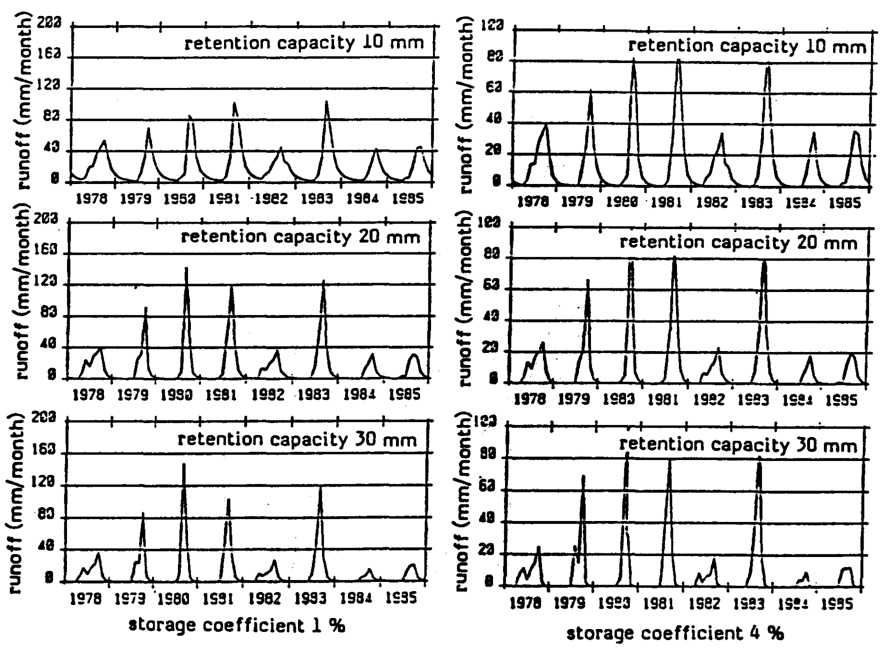

Fig. 24.6 Surface runoff simulated for three retention-capacity values and two storage coefficients

24.5 Data extension

24.5.1 Uniqueness of calibration

When it is seen that calibration has no unique solution, i.e. that various sets of parameters allow equally satisfactory simulation for a short period of observation (1978-1985 in this case), it is tempting to assume that a unique solution could be obtained for a longer period. This hypothesis effectively presumes that a long observation period (at least four or five times the half-recession time) should integrate years of very high and very low rainfall, together with exceptional rainfall sequences. The parameters must in this case cover the entire range: the level of soil retention must be nil at some periods and saturated at others, and the aquifer level must reach sufficiently low readings to identify recession and base level.

In order to check the hypothesis, the entire observation period (1959-1985) for daily rainfall in Ouagadougou was used, and evolution in the aquifer level was simulated using the various sets of parameters identified. Results are given in Fig. 24.7, which shows that, for the 17 year period, the levels calculated using the the various sets of parameters are virtually identical. It is therefore concluded that a long observation period does not improve the possibility of establishing a single set of parameters or a single recharge value.

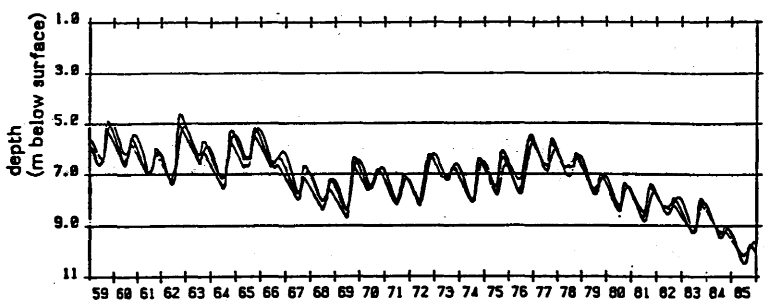

Fig. 24.7 Simulation of the 1959-1985 period for the three retention-capacity values and a storage coefficient of 1%

For purposes of a well-defined calibration, it is therefore believed that a sequence of piezometric observations is insufficient, even when the value of the storage coefficient is known. It is effectively necessary also to have access to "direct" and "delayed" runoff measurements and/or to measurements of the capillarity and water content in the unsaturated zone. This will provide all the elements in the balance, i.e.:

(1) runoff (measured)

(2) infiltration and evapotranspiration based on measurements in the unsaturated zone

(3) variation in the aquifer level (measured)

24.5.2 Reliability of the model

The model is calibrated on the eight years between 1978 and 1985. Before being used to extend data, it is necessary to check the model's behaviour by extrapolation (see Thiery 1988a, for instance). For three retention capacities (0, 20 and 40 mm), calibration for the first four years was made, the levels for the last four years being calculated without modifying the calibration. The reverse operation was also undertaken (calibration based on the last four years being checked for the first four years). Fig. 24.8 shows that results are very satisfactory. Calibration for the 1978-1981 period competently forecasts water levels for 1982 to 1985, although the latter are clearly lower than those for the calibration period. Calibration for the 1982-1985 period also, although less satisfactorily, allows calculation of levels for 1978 to 1981.

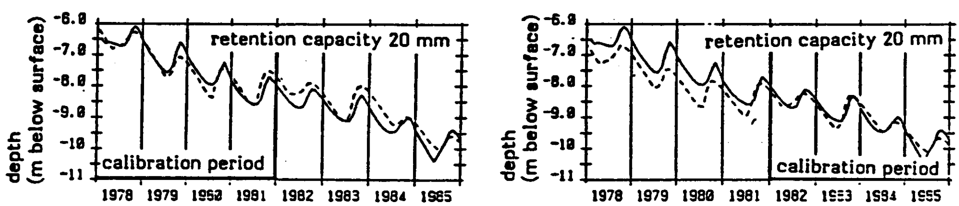

Fig. 24.8 Control of model calibration
(a) calibration based on the 1978-1981 period,
(b) calibration based on the 1982-1985 period

24.5.3 Simulation of return to more abundant rainfall

Analysis of annual recharge calculated for 1959-1985 shows that, for all sets of parameters, the 1981 recharge is very close to average recharge for the period as a whole. A sequence of ten years from 1985 was therefore simulated on the basis of the 1981 rainfall in order to observe the type and rate of aquifer reaction. Fig. 24.9, based on four hypotheses of calculation, shows a very slow rise, the aquifer taking more than ten years to reach a new equilibrium (on which seasonal fluctuation is superimposed). A single average year like 1981 will only produce a slight rise in water level after one year, and will not alone recharge the aquifer for several years, as might be assumed.

In order to place the period of study (1978-1985) in its context, the entire rainfall sequence from 1929 to 1985 was used. The model was accordingly readjusted for monthly rainfall figures (the only data available for 1929-1958), and very satisfactory calibration was obtained, with a correlation coefficient of 0.96.

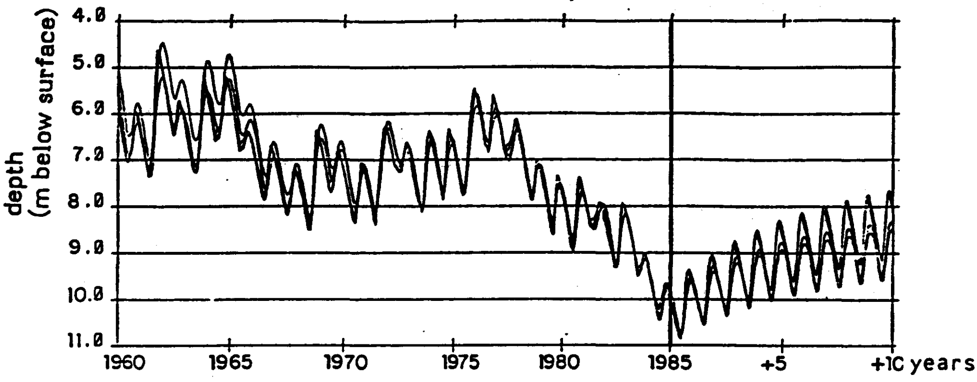

Fig. 24.9 Simulation of a period of ten average years (identical to 1981) from 1985

The entire period was then simulated. In order to show the effect of a return to more abundant rainfall, it was assumed that the 57 year sequence from 1929 to 1985 will be repeated from 1985. Fig. 24.10 shows that levels for the 1978-1985 period are the lowest since 1929, and that, if a wetter sequence occurs after 1985, a rise in water level to an average value will occur over a period of 7 to 10 years.

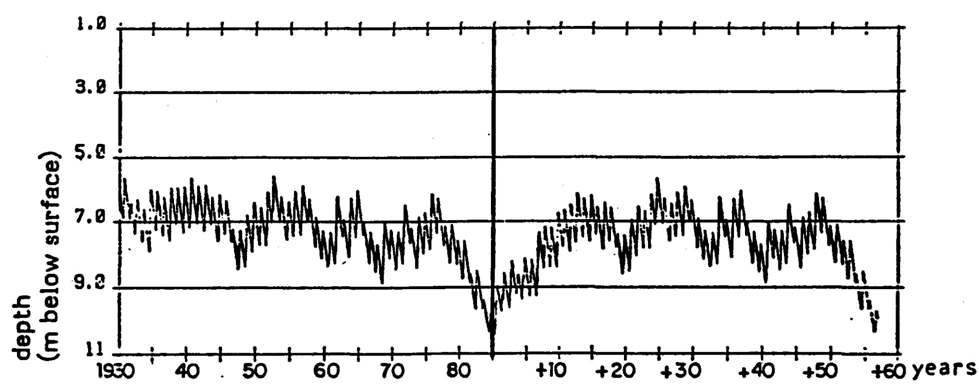

Fig. 24.10 Simulation of the 1929-1985 period prolonged by a 57 year rainfall sequence identical to that for 1929-1985

24.6 Conclusions

It is shown that a very simple lumped-parameter hydrological model of rainfall and water levels allows the correct reproduction of a piezometric sequence in semi-arid climatic conditions. For a detailed evaluation of the recharge, it is necessary to have access, in addition to a piezometric sequence (even if a long-duration one), to the storage coefficient and to a sequence of measurements for runoff (direct and delayed) or for capillarity and humidity in the unsaturated zone.

The various calibrations made show that the water retention capacity which allows optimum simulation of the piezometric sequence is between 10 and 30 mm. Assuming that the storage coefficient is equal to 1%, the computed annual recharge for the period 1978 to 1985 is estimated at between 23 and 45 mm/y. Despite the fact that calibration has no unique solution, a single extension of aquifer levels is obtained on the basis of a long sequence of rainfall data. The model thus allows extrapolation of levels for 1929-1977 based on rainfall figures. It shows that levels for the observation period (1978-1985) are the lowest in the 57 year period of historical rainfall records, and can easily be explained by natural climatic variation.

Two scenarios were simulated for the period after 1985. The first assumes a succession of years identical to 1981 (in which recharge was close to the average for the entire period). The second assumes repetition of the 1929-1985 sequence. Both scenarios show that the aquifer, which is not at this point in time in a critical state, should rise regularly and return to an average level in 7 to 10 years time.

It is thought that when water content observations are also available a semi-lumped parameter model (Thiery, 1988b) could help to obtain a unique calibration. When suction records in the unsaturated zone are available, a hydrodynamic model of the unsaturated zone (Thiery, 1988c) should give more accurate results.

25 RECOMMENDED BOOKS

This guidebook cannot hope to include all the information needed to understand hydrology, hydrogeology and recharge estimation. Nor can it detail all the techniques required, especially those outside the direct field of recharge. Therefore the authors have compiled the following list of recommended texts. Wherever possible, books and manuals from international agencies have been selected, as these are often available free or cheaply in developing countries. Addresses of the relevant agencies are also included. Widely available commercial literature has been selected if possible. Those marked * are especially recommended in relation to groundwater recharge studies.

25.1 General hydrogeology

S.N. DAVIS & R.J.M. DE WIEST, 1966. *Hydrogeology*. John Wiley & Sons, New York, Chicester.

* R.A. FREEZE & J.A. CHERRY, 1979. *Groundwater*. Prentice-Hall, Englewood Cliffs, New Jersey.

C.W. FETTER, 1988. *Applied hydrogeology* (2nd edition). Charles E Merrill Publishing Co., Columbus, Ohio.

D.K. TODD, 1980. *Groundwater hydrology* (2nd edition). John Wiley & Sons, New York, Chicester.

25.2 Groundwater hydraulics

H. BOUWER, 1978. *Groundwater hydrology*. MacGraw-Hill.

G.P. KRUSEMAN & N.A. DE RIDDER, 1981. *Analysis and evaluation of pumping test data*. Bulletin 11, ILRI, Wageningen.

A. VERRUYT, 1982. *Theory of groundwater flow* (2nd edition). MacMillan, London.

25.3 Groundwater modelling

J. BOONSTRA & N.A. DE RIDDER, 1981. *Numerical modelling of groundwater basins*. ILRI publication 29, ILRI, Wageningen.

J.D. BREDEHOEFT (ed), 1982. *Groundwater models*. Studies and Reports in Hydrology 34, UNESCO, Paris.

K.R. RUSHTON & S.C. REDSHAW, 1979. *Seepage and groundwater flow*. John Wiley & Sons, New York, Chicester.

25.4 Hydrology and hydrometry

* FOOD AND AGRICULTURE ORGANISATION, 1981. *Arid zone hydrology*. Irrigation and drainage paper 27, FAO, Rome.

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R.W. HERSCHY (ed), 1978. Hydrometry - principles and practices. John Wiley & Sons, New York, Chicester.

25.5 Hydrogeological provinces

A. BURGER & L. DUBERTRET (eds), 1984. *Hydrogeology of karstic terrains: case histories*. International contributions to hydrogeology, vol. 1, Heise, Hannover. (Mixed English and French.)

P.E. LaMOREAUX (ed), 1986. *Hydrology of limestone terranes: Annotated bibliography of carbonate rocks 3.* International contributions to hydrogeology, vol. 2, Heise, Hannover.

P.E. LaMOREAUX, B.M. WILSON & B.A. MEMON (eds), 1984. *Guide to the hydrology of carbonate rocks.* Studies and Reports in Hydrology 41, UNESCO, Paris.

I. LARSSON (ed), 1984. *Groundwater in hard rocks.* Studies and Reports in Hydrology 33, UNESCO, Paris.

25.6 Recharge

* I. SIMMERS (ed), 1988. *Estimation of natural groundwater recharge.* NATO ASI Series C, Vol. 222 (Proc. of the NATO advanced research workshop, Antalya, Turkey, March 1987.) D. Reidel Publ. Co., Dordrecht.

WATER RESOURCES RESEARCH CENTER, 1980. *Regional recharge research for southwest alluvial basins.* Final report on USGS contract 14-08-0001-18257. Dept of Hydrology and Water Resources, University of Arizona, Tuscon, Arizona 85721, USA.

25.7 Isotope hydrology

* INTERNATIONAL ATOMIC ENERGY AGENCY, 1983. *Guidebook on nuclear techniques in hydrology.* Technical Report Series, no. 91, IAEA, Vienna.

25.8 Addresses

FAO: Distribution and Sales section, FAO, Via delle Terme di Caracalla, 00100 Rome, Italy.

Heise: Verlag Heinz Heise, PO Box 2746, D-3000 Hannover 1, Federal Republic of Germany.

IAEA: Division of publications, IAEA, Wagramerstrasse 5, PO Box 100, A-1400 Vienna, Austria.

ILRI: International Institute for Land Reclamation and Improvement, PO Box 45, 6700 AA Wageningen, The Netherlands.

UNESCO: Division of Water Sciences, UNESCO, 7 Place de Fontenoy, 75700 Paris, France.

WMO: Publication and Sales Unit, WMO, PO Box 5, 1211 Geneva, Switzerland.

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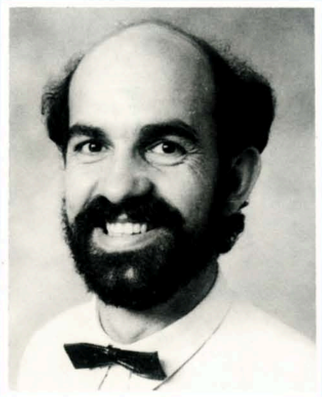

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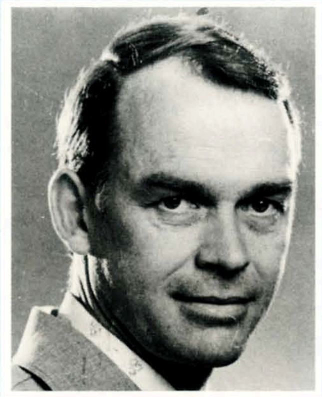

Ian Simmers

This manual of practice, prepared by an IAH workgroup as a contribution to the International Hydrological Programme od Unesco, especially refers to the world's arid and semi-arid zones, these being the areas where the need for reliable estimates of groundwater recharge are greatest.

Results of a workshop held in Turkey in 1987 on estimation of natural and man-induced groundwater recharge are also included.

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