8 WATER BALANCE

Once the measurements of the components of the hydrological cycle are complete, the analysis of the hydrology of a basin or a region is best approached through the water balance. This simply expresses the fact that, over any time period, water input to an area must equal output and changes in storage. Input is likely to be mainly rainfall but may include river inflow or groundwater inflow; output will include evaporation, river outflow and groundwater outflow, while storage changes include increases in soil moisture and groundwater storage. Examples of water balance analysis may take the form of balance sheets, but may also include conceptual models which express physical understanding of the links between different processes. It will be seen from examples that the ease of water balance study depends on the type of climate of the area.

TYPES OF CLIMATE

The first step in examining the climate of an area is to compare typical values of average monthly rainfall and potential transpiration or evaporation. Rainfall averages are usually available in standard meteorological yearbooks, but evaporation measurements or estimates are not so readily available as standard statistics. It may be necessary to derive potential transpiration from standard meteorological records using the Penman technique (Chapter 4). In fact climate data are available for this purpose (FAO, 1993). Examples are compared in Fig. 4.1 of average rainfall and potential transpiration for a humid area, a semiarid area and a seasonal monsoon climate. The humid example is from Nuwara Eliya in the central highlands of Sri Lanka, where the incidence of rainfall from both the northeast and southwest monsoons provides a plentiful supply of moisture in most months, and average rainfall exceeds potential transpiration (Fig. 4.1(a)) to give a surplus throughout the year. In these circumstances water balance studies are relatively simple, as the soil moisture deficit seldom exceeds the root constant (Chapter 5) when the transpiration is assumed to fall below the potential rate. In fact runoff can often be estimated as the difference between rainfall and potential transpiration.

By contrast, the comparison between average rainfall and potential transpiration for Francistown in northeast Botswana (Fig. 4.1(b)) reveals that there is no month with a surplus on average. The relatively small depth of annual runoff, typically 10–50 mm, is caused by short periods of intense rainfall in storms which are often isolated and variable. Comparisons between rainfall and runoff must either be based on a comprehensive network of recording raingauges to compare areal rainfall intensity and soil infiltration rates, or be based on time and areal averages and thus remain imprecise. In these semiarid areas estimates of runoff from measured rainfall tend to be empirical, and the runoff process is sensitive to differences in basin topography and geology because the climate leads to low runoff coefficients.

In wide areas of India and Africa there is seasonal variation between arid and humid conditions, and in these conditions the importance of water resources management is evident. These conditions are illustrated (Fig. 4.1(c)) by monthly rainfall and transpiration records for Bhopal in central India. Although the climate appears to be more complex than the other examples, it is possible to simplify the situation by treating the monsoon period as a period of surplus, and the prolonged dry season as a period of deficit with moisture dependent on soil moisture storage.

HUMID WATER BALANCE

The water balance approach is most easily applied in areas of humid climate, where the transpiration can be assumed to be close to the potential rate. Uncertainty arises when the soil moisture deficit increases to the root constant appropriate to the dominant vegetation, when the soil moisture is not readily available and the transpiration falls below the potential rate. Thus the runoff from a humid basin or region should correspond to rainfall less potential transpiration, not just on an average basis but also from year to year. The example of the Tongariro area in New Zealand has already been mentioned in Chapter 3, and is discussed further in Chapter 9. The difference between rainfall and runoff was uniform, so that the flow series could be extended quite simply.

Sri Lanka

A classic case of a water balance study in a humid environment involved the Mahaweli Ganga basin in central Sri Lanka. The Victoria dam project (Piper *et al.*, 1994) required an appraisal of the reliability of the river flow records of the area, and also an extension of flow records to ungauged sites and a common period. The network of long-term raingauges in this region was excellent, as the development of tea estates during the 19th century had led to an interest in rainfall distribution and about 100 stations were available around the central highlands. An annual isohyetal map (Fig. 8.1) was derived from these records and showed that the average rainfall ranges from over 5000 mm on the westerly slopes of the mountains to less than 2000 mm in areas of rainfall shadow which are exposed to neither monsoon direction. The seasonal distribution of rainfall varies with aspect and thus exposure to the southwest and northeast monsoons, which in turn provide rainfall from May to August and from October to January. Other meteorological processes contribute rainfall during other seasons. Although the seasonal distribution over the area as a whole reflects the relative contributions of the two monsoons, the upper Mahaweli basin which feeds the Victoria dam is reasonably homogeneous and rainfall is sufficient to maintain a surplus throughout the year. Even in the tributaries where rainfall is more seasonal, the soil moisture deficit is unlikely to reach a value which would reduce transpiration significantly.

The Penman approach was used to estimate open water evaporation and potential transpiration from meteorological records at three sites in the project area, and these estimates showed a decrease of annual totals with elevation. Flow records were available at a number of sites in the area. A comparison of mean annual rainfall and runoff (Fig. 8.2) showed, as expected, a reasonably constant difference between basin rainfall and runoff depth, which in turn corresponded with potential transpiration. However, this comparison drew attention to the fact that at some sites the runoff records were low, and the flows at these sites were reviewed.

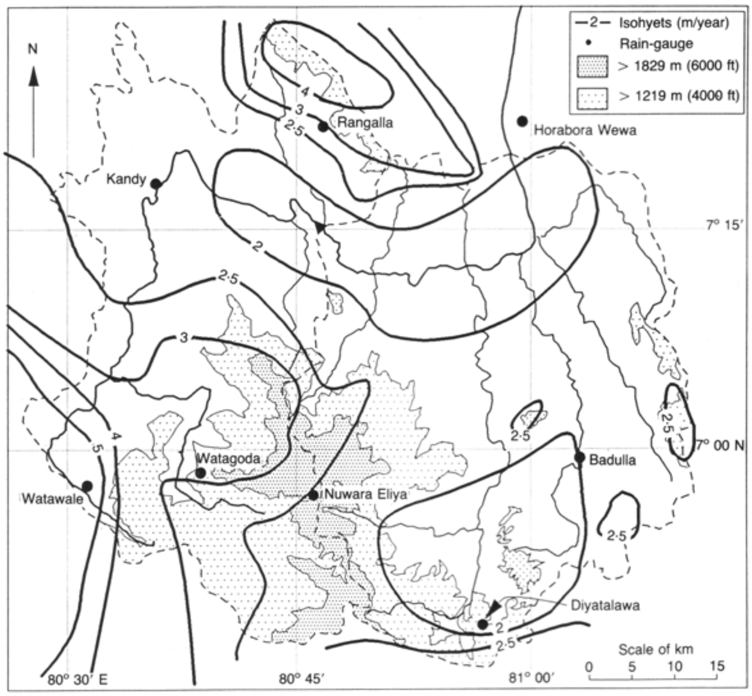

Fig. 8.1 *Central Sri Lanka: annual isohyetal map, m (1931–1960) (after Piper et al., 1994).*

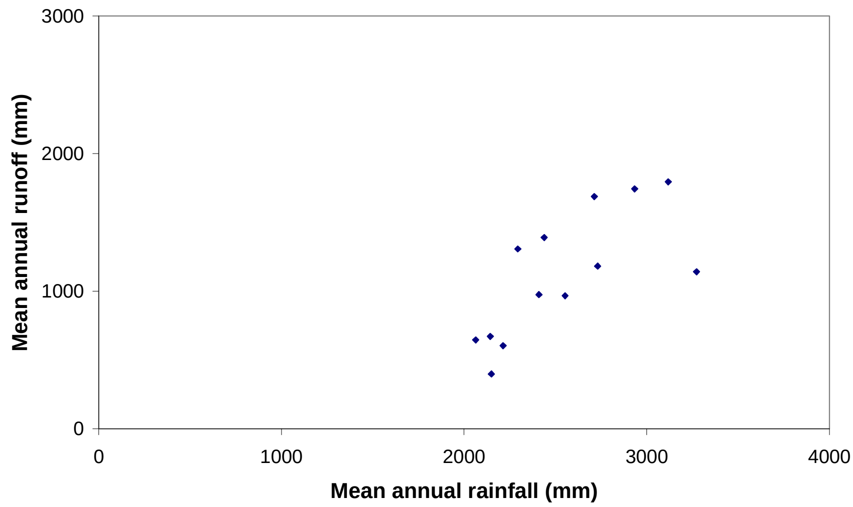

Fig. 8.2 *Sri Lanka: mean annual rainfall and runoff at Mahaweli basin stations (mm).*

There is a similarity between comparisons of mean annual rainfall and runoff at a number of sites in a region, and comparisons of successive annual rainfall totals and runoff depths at a single station. The available records had been checked internally by comparison with Thiessen-based estimates of basin monthly rainfall from selected stations. This information was also used in a comparison of annual rainfall and runoff

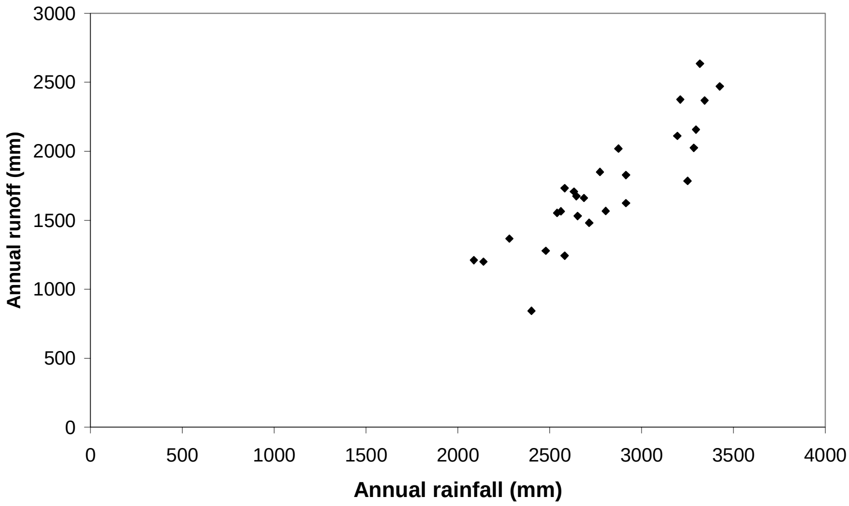

Fig. 8.3 *Sri Lanka: annual rainfall and runoff of the Mahaweli Ganga at Gurudeniya, 1944–1970 (after Piper et al., 1994).*

at all sites, and this comparison confirmed the reliability of most stations. An example (Fig. 8.3) compares the annual flows of the Mahaweli Ganga at Gurudeniya and the basin rainfall series. The correlation between the rainfall and runoff series confirms the reasonable validity of the flow records.

This finding made it possible to estimate mean runoff at key ungauged sites by subtracting potential transpiration from basin rainfall averages derived from the isohyetal map. Consistent series for reservoir operation studies were derived from a statistical comparison between short-term and long-term stations, taking account of cross-correlation and serial correlation within the records. The study was made both simpler and more reliable by the homogeneous and humid nature of the region.

Sarajevo region, Bosnia

In a study in 1998 of the water resources of the Sarajevo region, another example of a humid water balance made it relatively easy to obtain consistent flow records for the area. Sarajevo itself is surrounded by an area of mainly forested hills which are drained by a number of streams which feed into the Bosna River which in turn flows into the Danube. Sarajevo city lies in a valley where a number of these streams converge, and it lies in an area of relatively low rainfall. The monthly average rainfall is compared in Fig. 8.4 with estimates of potential transpiration, derived from Penman estimates of open water evaporation using a factor of 0.8.

It is clear from comparison of Sarajevo rainfall records with those of stations to the north and south that the area receives a combination of Mediterranean winter rainfall with a summer maximum typical of a Central European climate. Although individual years are subject to variations from the average in both sources of rainfall, the average seasonal distribution is extremely uniform. The potential transpiration, on the other hand, is highly seasonal with its link to net radiation and thus temperature. The result is a winter surplus of precipitation, much of which in fact occurs as snowfall, and a relatively small summer deficit.

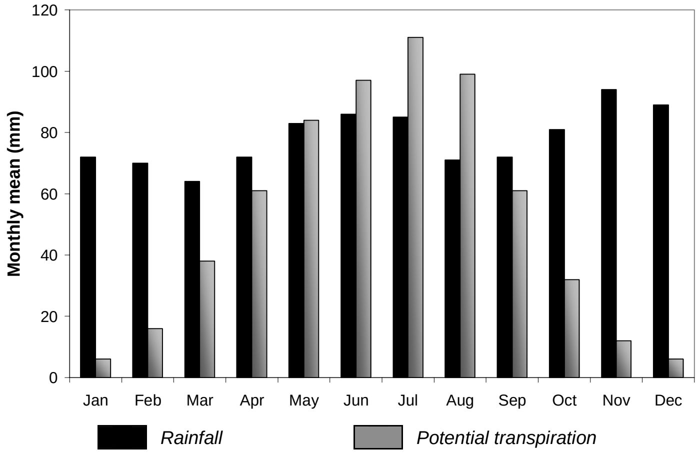

Fig. 8.4 *Sarajevo: mean monthly rainfall and potential transpiration.*

Plate 8.1 *Woodland east (upstream) of Sarajevo.*

It was evident from a visit to the basin headwaters, which enjoy higher rainfall and contribute the bulk of the river flow, that even towards the end of the summer, when the soil moisture deficit would be at its highest, the woodland of the basins (Plate 8.1) appeared to be transpiring freely. It could therefore be assumed that the rainfall was sufficient for the soil moisture deficit rarely if ever to exceed the root constant of the woodland, and therefore that the area could be treated as a humid area. A simple monthly model was developed, with a small fraction (0.1) of the gross precipitation leading to direct runoff, and the balance of the net rainfall contributing to the soil moisture store of 300 mm capacity. This was assumed to transpire at the potential rate and provide runoff when the store becomes full. The runoff was assumed to reach the gauging station over a period of three months; the need for three monthly recession constants to total unity defines them as 0.54, 0.30 and 0.16. An example of this modelling of the River Miljacka at Sarajevo is illustrated in Fig. 8.5. The basin average

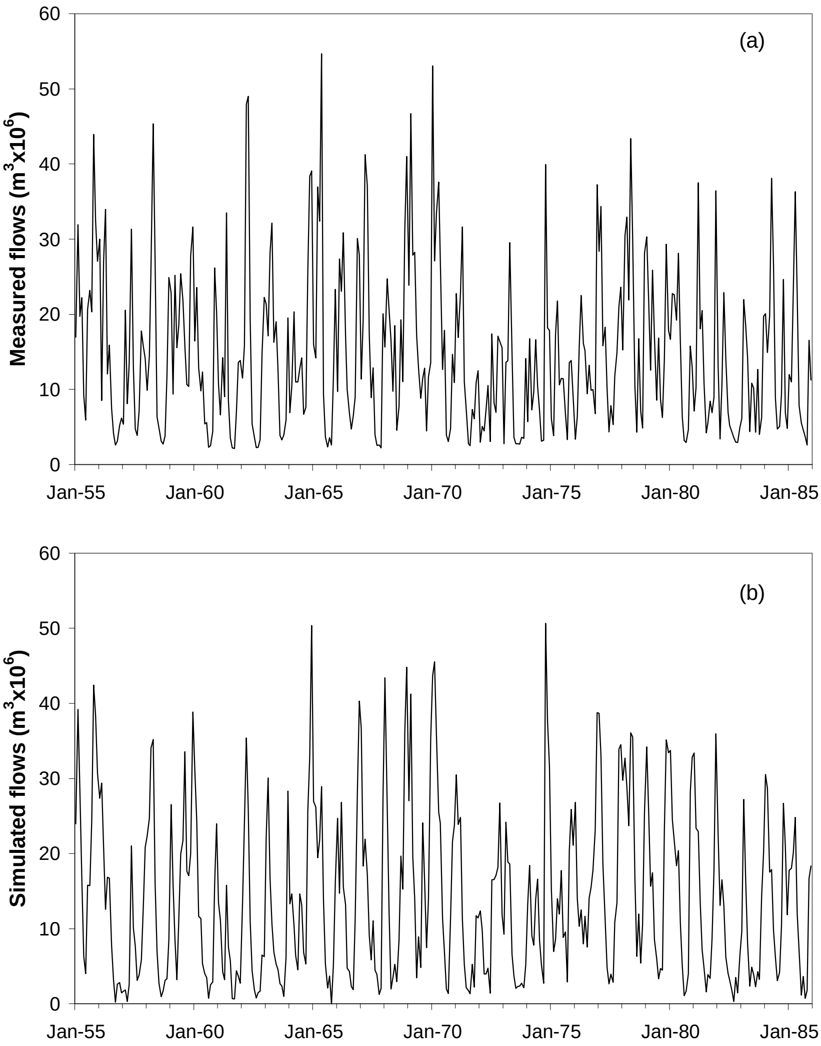

Fig. 8.5 *Miljacka at Sarajevo: (a) measured, and (b) simulated, monthly runoff.*

precipitation was estimated from an isohyetal map as 1200 mm and its monthly distribution from four related rainfall stations using the isopercentile method. This simple model provided reasonable comparisons with measured flows both as monthly series and in a cumulative comparison. The main difference between the measured (Fig. 8.5(a)) and simulated flow series (Fig. 8.5(b)) was a lag in the winter and spring flows which could be explained by snow storage which had not been taken into account in the initial modelling.

This model was then used to generate flow series at ungauged sites, and also to extend all flow series to a standard period in order to represent the flow series over the whole of the contributing area. Again the success of the simple model could be attributed to the nature of the humid water balance.

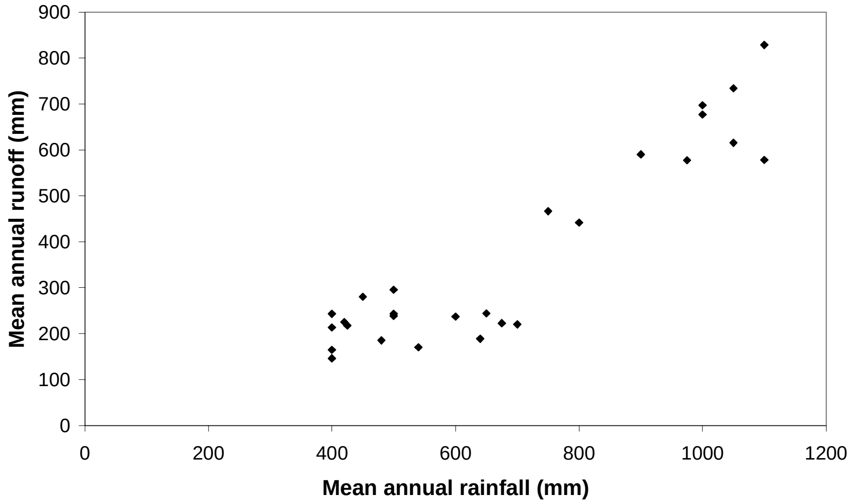

Fig. 8.6 *Northern Turkey: mean annual precipitation and runoff.*

Other humid basins

Areas like upland Britain are evidently humid, as the rainfall in most parts is usually adequate to maintain transpiration at the potential rate through most of the year, and it is rare that the summer has a severe drought during which soil moisture storage cannot maintain transpiration. Most of the runoff occurs during the winter season, but the transpiration occurs at the potential rate throughout the year.

However, there are regions where the maximum rainfall coincides with a season of low potential transpiration. These areas are effectively humid because the runoff derives from a seasonal surplus and the basin loss is relatively constant. For example, the precipitation distribution over the region of northern Turkey inland from the Black Sea is concentrated in the winter months when temperatures and potential transpiration are both low. The result is that runoff occurs between December and April, and is higher than would be expected from the annual precipitation. Figure 8.6 illustrates the mean annual precipitation and runoff for a number of basins. The runoff corresponding to basins of 500 mm precipitation is about 200 mm, while for basins with precipitation of 1000–1100 mm the runoff is as high as 600–700 mm. This is much higher than the runoff in other parts of the world with similar precipitation, for example Botswana and Yemen described in the next section, because of the low potential transpiration during the winter precipitation season.

ARID WATER BALANCE

Water balance analysis is less simple in arid or semiarid areas, mainly because actual transpiration is much less easy to predict from potential transpiration. Because surplus soil moisture seldom gives rise to widespread runoff, the runoff process is more likely to depend on surface flow from local intense storms when rainfall exceeds local infiltration. This is of course the classical Hortonian concept of runoff, based on United States experience, which has not been entirely convincing to hydrologists working in more temperate conditions, who may never have observed surface runoff.

Such conditions require more complex analysis than the humid situation described previously. If the runoff process were to be modelled physically, it would be necessary to monitor rainfall on a very detailed scale, both in terms of the time scale required to estimate rainfall intensity, and also to take account of the greater spatial variability in shorter time intervals. However, it will be shown that it is possible to use models based on monthly time intervals by making assumptions about the distribution of rainfall and basin conditions.

Northeast Botswana

Successive studies in Botswana in 1968, 1976 and 1986 (Parks & Sutcliffe, 1987) illustrate the problems which semiarid climates present. Average annual rainfall decreases from over 500 mm north of Francistown to less than 350 mm in an area of low rainfall in the lower Motloutse basin to the south (Fig. 8.7). The rainfall season is from November to March, when the potential transpiration is at its highest. Monthly averages for Francistown (Fig. 4.1(b)) show that annual rainfall is less than a third of transpiration and its seasonal cycle is similar. Thus potential transpiration exceeds average rainfall in all months, and runoff mainly results from local convective storms and high rainfall intensity. Annual runoff is low at a basin average of 10 to 50 mm.

In fact it was only after 20 years of fairly regular visits that it was possible to observe significant runoff for the first time (compare Plates 8.2 and 8.3). During heavy

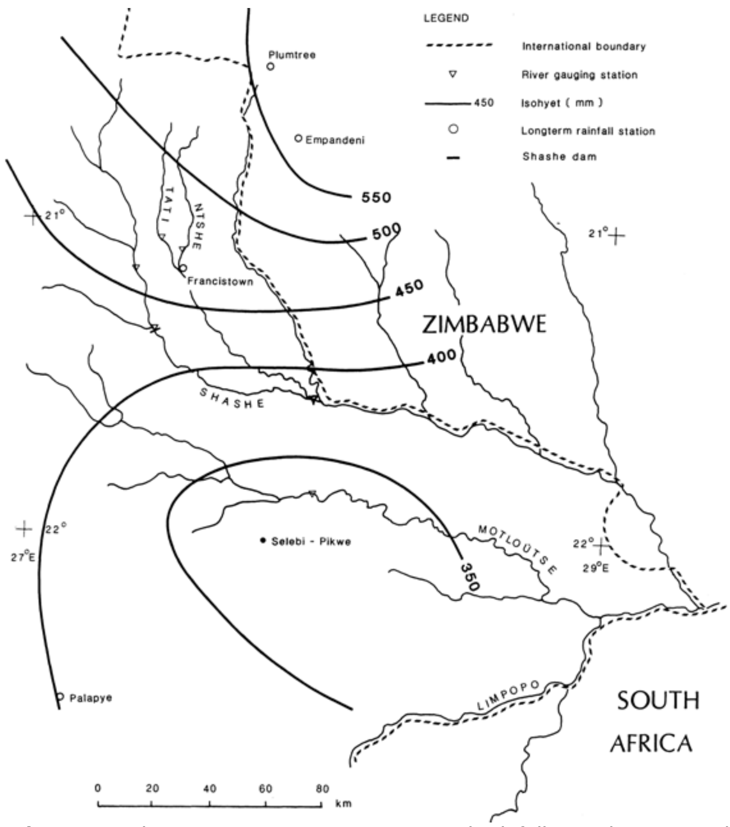

Fig. 8.7 *Northeast Botswana: average annual rainfall, mm (1939–1969) (from Parks & Sutcliffe, 1987).*

Plate 8.2 *River Shashe gauging weir near Francistown, Botswana.*

Plate 8.3 *Shashe weir during spate, 1986.*

storms runoff appears to occur over parts of the basin where rainfall intensity exceeds infiltration rates on less permeable soils, bringing into operation classical hydrological concepts like depression storage and surface runoff. The runoff generated locally forms sheet flow and is concentrated in gullies and small streams and thus into river flows.

The headwaters of the Shashe tributaries north of Francistown are on Basement Complex rocks with moderate relief and soil depth, and gullies develop into sandy tributaries with rock outcrops and finally into broad sand-bed rivers draining to the Limpopo. To the southwest the Kalahari Beds provide deep sand cover, especially in the headwaters of the Motloutse (Fig. 8.7). River flows occur in a small number of spates, but leave the sandy river beds apparently dry for much of the time, even in the wet season. As a result flow measurement is extremely difficult, but a combination of gauging weirs on rock outcrops (Plate 8.2) in the upper tributaries and rated sections in the lower reaches have over the years provided records at key sites. The number of significant streams is limited, and the locations of gauging sites were partly determined by early investigations of potential dam sites.

In these conditions, water balance studies are likely to be largely empirical, as physical modelling would require a dense network of measurements. It is possible to compare average runoff depth at gauging sites with mean basin rainfall, taking note of other factors such as geology, soils, topography and vegetation. Alternatively statistical or semi-empirical conceptual models may be used to extend flow records where sufficient rainfall and runoff records are available for calibration, or to generate approximate flow series at ungauged sites.

In early investigations in 1968, only three flow records each of six years' duration were available, and a detailed isohyetal map had to be compiled as a basic tool of the study. Use was made of regional studies in southern Africa by Midgley (1952), which suggested that although mean runoff depth depends on a number of factors like rainfall, relief, geology, soils and vegetation, these factors are to a large extent linked and relations between mean rainfall and runoff depth can be determined in terms of regional vegetation and land use. He provided a table linking mean runoff with rainfall in zones classified by vegetation and land use, based in part on flow records on Limpopo tributaries.

The mean runoff at the three gauged sites was deduced by comparing rainfall during the period of flow records with the long-term mean, and the results were compared with Midgley's table. However, the sensitivity of the choice of zone made it risky to estimate the runoff at ungauged sites, and a reservoir site on the Shashe below the gauging station was selected for detailed study. A simple relation between monthly rainfall and runoff was derived and used to deduce the probable reservoir level at the start of the critical drought. This study drew attention to the importance of measured flows in the area, and gauging was extended to other potential reservoir sites.

A later regional reconnaissance study of potential reservoir sites in eastern Botswana was carried out in 1976. After likely dam sites had been identified from air photography, estimates of mean annual runoff and its variability were required at 45 sites in northeast Botswana. Flow records at eight sites in northeast Botswana were adjusted to a common period and runoff depths were compared with mean basin rainfall, together with 10 records from adjacent Shashe tributaries in Zimbabwe. On the assumption that land use and slopes were reasonably homogeneous, a relationship

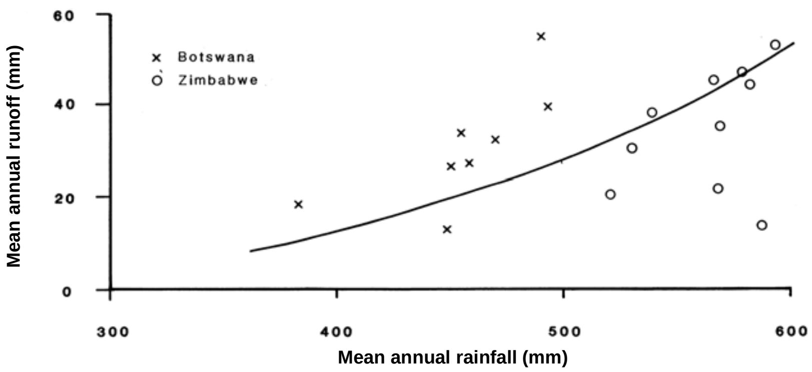

Fig. 8.8 Northeast Botswana: mean annual rainfall and runoff (mm) (from Parks & Sutcliffe, 1987).

between runoff and rainfall (Fig. 8.8) was used to make preliminary estimates of mean runoff at reservoir sites, from which a smaller number were selected for further study. The variability of annual flows was deduced from dimensionless cumulative flow frequency curves for durations from 1 to 7 years derived from adjacent South African flow records (Midgley & Pitman, 1969). These were tested using a rainfall–runoff model of the Shashe flows.

By 1986, when a third study was made, these and other studies had resulted in the installation of gauging stations near the most promising reservoir sites. Measured flows were available for periods ranging from 15 to 23 years. This period was thought sufficient to fit a rainfall–runoff model to extend the flows over the period of long-term rainfall records. Rainfall records at Francistown and at two other sites with records from 1922, were supplemented by adjacent stations in Zimbabwe, some of which dated from 1899. The isopercentile method (Chapter 3) was used with the basin averages to derive long-term rainfall series for each gauged basin.

Pitman model

The Pitman model (1973) was fitted to the observed rainfall and runoff series for each basin. This model (Fig. 8.9) was developed to represent hydrological processes over a variety of climates in South Africa, and was initially designed to convert monthly rainfall depths and potential transpiration data to monthly river flows. It is semi-empirical in that the rainfall–runoff process is simplified and expressed in semi-statistical relations. The model takes into account interception storage, the impervious fraction and a range of infiltration rates, soil moisture storage and basin lag. The model uses 12 parameters, some of which are important in humid areas and others in semi-arid areas. Monthly rainfall is converted to a synthetic symmetrical S-shaped curve, in which the deviation from a uniform rainfall rate is related to monthly totals. Monthly interception is deduced from monthly rainfall as a function of interception storage. Runoff in semiarid areas is related to the impervious fraction and the variation of infiltration rates over the basin, and is modelled by a symmetrical triangular frequency distribution between the minimum and maximum infiltration rates.

Where runoff depends on water balance rather than infiltration rates, basin evaporation is assumed to decrease below the potential rate as soil moisture storage decreases below the maximum, while the runoff coefficient is linked to soil moisture by a power relation with a threshold soil moisture and a runoff coefficient at maximum storage. Runoff is lagged by treating the basin as a linear reservoir. Although the full model has 12 parameters, only five were important in these semiarid conditions: the maximum soil moisture capacity, the minimum and maximum catchment absorption rate, the interception storage, and the runoff lag.

The fixing of the model parameters is based on comparison of simulated and observed flows. Although several of the parameters have hydrological titles, it is recommended (Pitman, 1973) that their estimation should be adjusted by fitting the modelled to the observed runoff. The calibration of the model was based on comparisons of simulated and observed mean annual flow, the mean and standard deviation of the logarithms of annual flows, and the seasonal distribution of runoff. Advice is given by Pitman on the effect of changes in the model parameters; once the parameters had been adjusted by recommended methods, the Botswana runoff series were extended back to 1922 using long-term rainfall series. However, the lack of close correlation

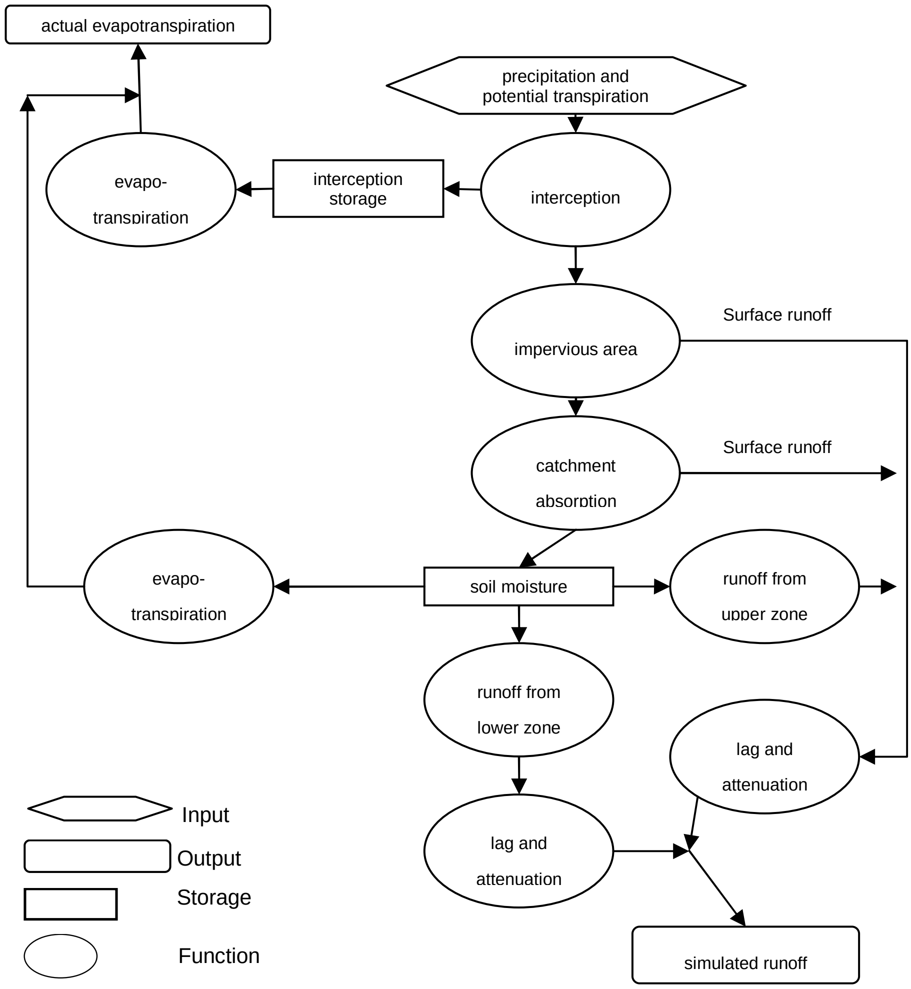

Fig. 8.9 *Flow diagram of the Pitman model (after Pitman, 1973).*

between simulated and observed runoff during the calibration period (Fig. 8.10) shows that the natural variability of the runoff process (Plate 8.3) in this semiarid environment makes it difficult to model the rainfall–runoff process accurately in this area.

As the available flow records for this region have increased in number and length, they have enabled more comprehensive methods of analysis to be used. Once a reasonable period of runoff records could be used, it was possible to use rainfall records to extend this period further. As a result of the increasing length of direct or extended flow record, the standard error of estimate of the mean annual runoff at the Shashe dam has decreased steadily, from 30 $m^3 \times 10^6$ in 1968 to 20 $m^3 \times 10^6$ in 1976 and 12 $m^3 \times 10^6$ in 1986. However, the area appears to be subject to fluctuating periods of dry and wet years (Fig. 8.11), and the successive estimates have varied from 86 $m^3 \times 10^6$, up to 99 $m^3 \times 10^6$ and down to 74 $m^3 \times 10^6$ over the years. In view of this variability of available water resources which occurs over the years in some climates, it is important to use as long a period of flow record as possible. This topic is discussed further in Chapter 9.

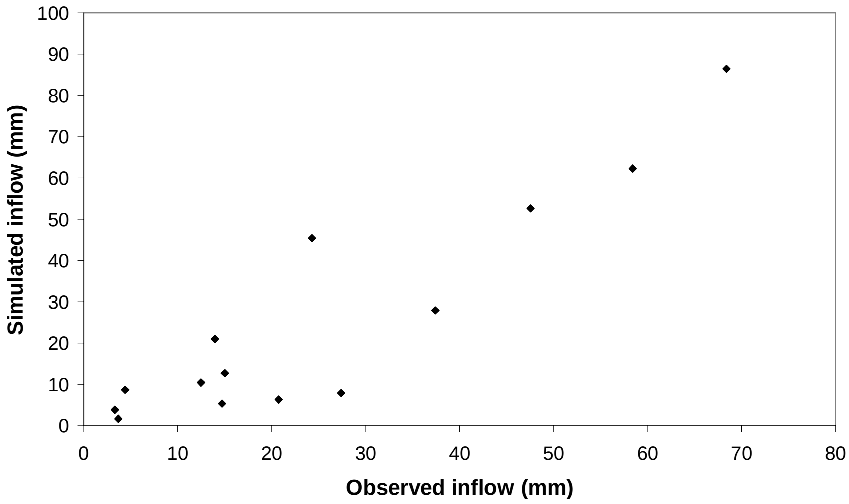

Fig. 8.10 *Shashe, northeast Botswana: measured and simulated annual runoff, 1970–1983 (after Sutcliffe, 1995).*

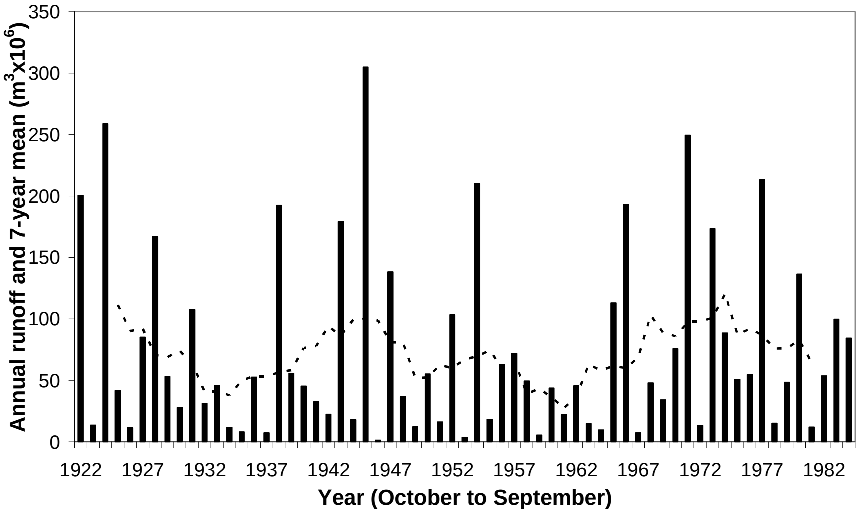

Fig. 8.11 *Shashe dam site: annual flows 1922–1984.*

Iran

Two examples from Iran illustrate the problems of deriving a water balance in mountainous areas. The first example, from the Zagros Mountains in western Iran (Sutcliffe & Carpenter, 1967), is mentioned in Chapter 3 as the estimation of mean precipitation over the different river basins required adjustment to a common elevation (Fig. 3.5) as precipitation varied with both elevation and position. Although the mean precipitation varied between 300 and 700 mm, the mean runoff depth varied between 40 and 300 mm. The open water evaporation, estimated by the Penman method, decreased from 2000 mm at an elevation of 150 m at Dezful, to 1600 mm at 1300 m at Kermanshah, but the seasonal distribution of precipitation was such that the surplus

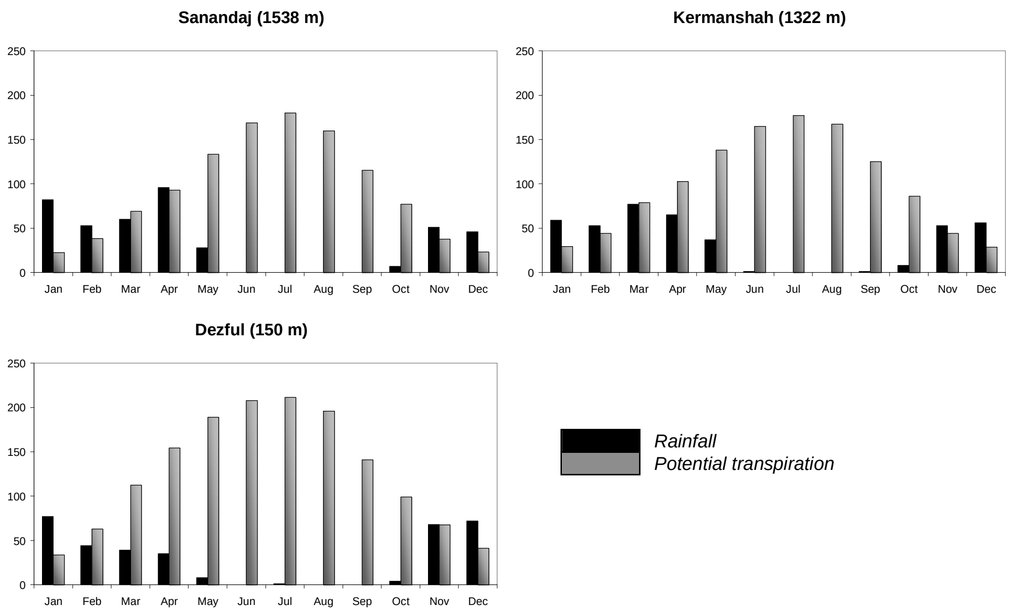

Fig. 8.12 *Kermanshah area, Iran: rainfall and potential transpiration at key sites, mm.*

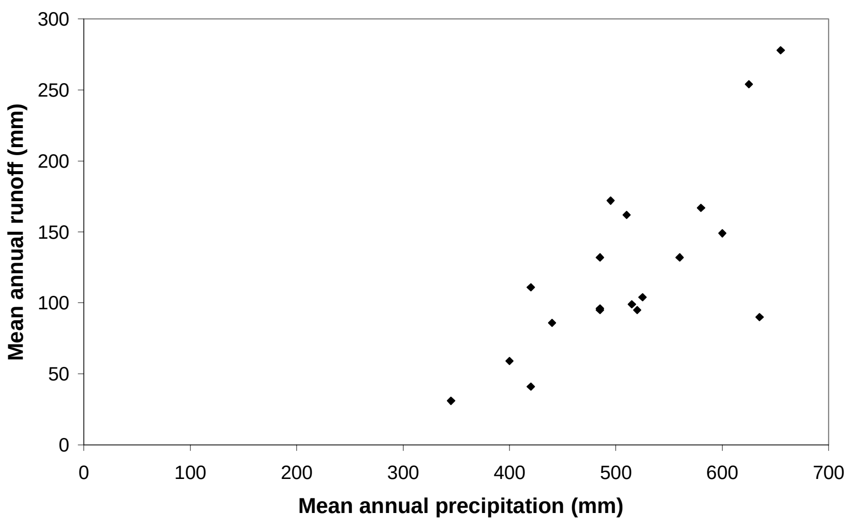

Fig. 8.13 *Zagros region, Iran: rainfall and runoff (after Sutcliffe & Carpenter, 1967).*

was relatively large. The seasonal distributions of precipitation and potential transpiration at key sites are illustrated in Fig. 8.12. These show that a small water surplus occurs when the winter precipitation exceeds the low transpiration, particularly at high elevations. On the other hand, the potential transpiration is high during the summer, when there is little or no rainfall. The relation between mean precipitation and runoff at sites where measured flows were available, after adjustment to a common period, is illustrated in Fig. 8.13. Further analysis of the records revealed that actual evaporation

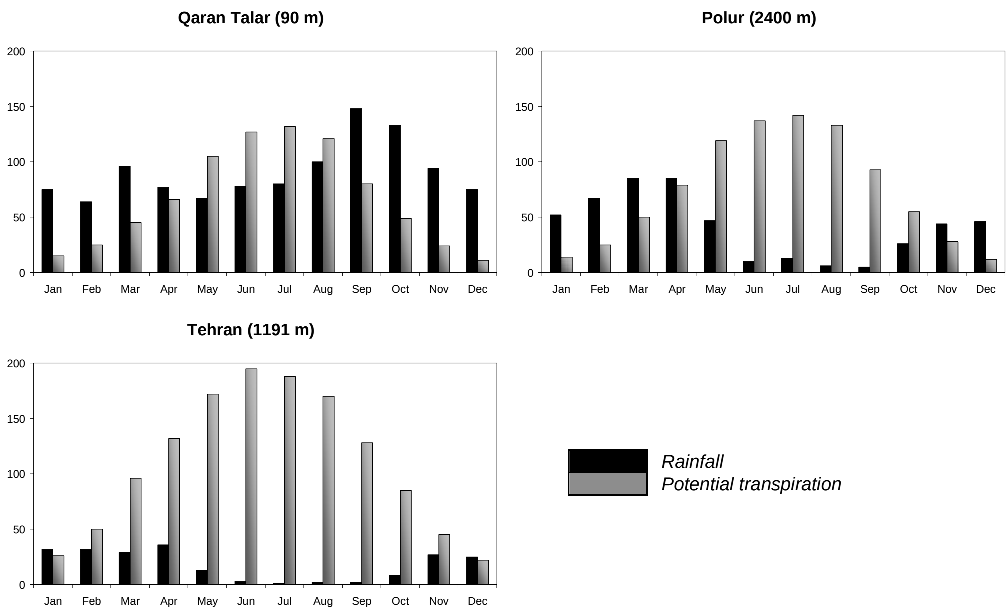

Fig. 8.14 *Alborz area, Iran: rainfall and potential transpiration at key sites, mm.*

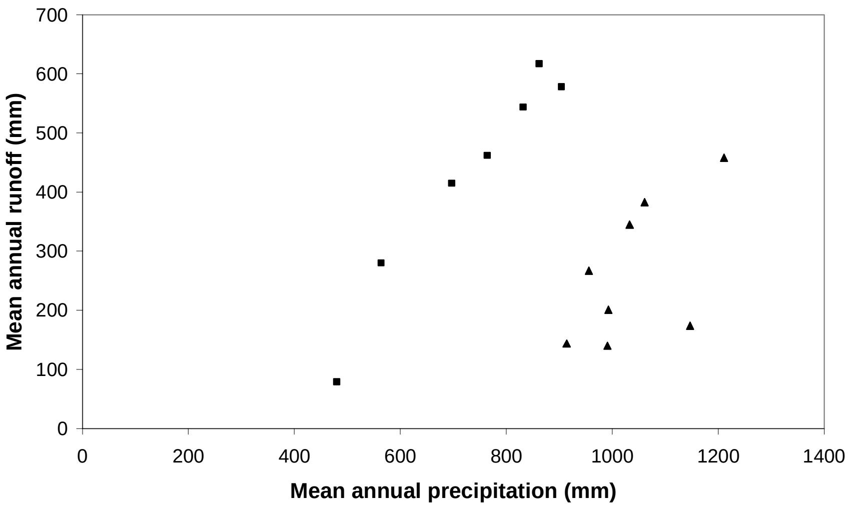

Fig. 8.15 *Alborz region, Iran: rainfall and runoff; ▲ northern and ■ southern basins.*

is quite conservative. Indeed, though precipitation depends greatly on elevation, it appears that the combined effect of lower potential transpiration and higher precipitation at higher altitudes produce a fairly constant actual evaporation loss. This fact made it possible to make reasonable estimates of mean runoff at ungauged sites.

A second example from Iran contrasts a humid region with an adjacent semiarid region separated solely by aspect (Sutcliffe & Swan, 1970). To the south of the Alborz massif, on either side of Tehran, are a number of basins with a water balance similar to the Zagros region in western Iran. The mean precipitation increases from 210 mm at Tehran at an elevation of 1200 m, to some 800 mm at elevations of 2500 m on the southern slopes, but the precipitation is confined to the winter months. To the north of the watershed, on the basins facing the Caspian Sea, the precipitation is higher, up to 1200 mm, with incidence throughout the year. Monthly average precipitation and potential transpiration at key stations to the north and south of the divide are compared in Fig. 8.14, which illustrates the contrast between the climate of the two aspects. This contrast was discussed in Chapter 3 and the isohyetal map was shown as Fig. 3.6. However, the even distribution of rainfall on the northern slopes has allowed forest to survive and become thick in some areas, and the measured runoff from the northern rivers is in fact slightly lower than from the drier but relatively bare basins to the south.

The mean runoff values from basins on either side of the mountains are compared in Fig. 8.15; the difference between precipitation and runoff, or the apparent loss, is about 750 mm on the northern slopes but only about 300 mm on the southern slopes. The difference in climate between the two slopes is responsible for the difference in vegetation, but the vegetation in turn is responsible for the higher transpiration and interception loss on the forested basins. Once the concept of soil moisture storage is included in the comparison of precipitation and transpiration, the effects of the different climates on runoff depth are explicable. This area provides an interesting comparison of a humid and a semiarid set of river basins separated simply by the Alborz watershed.

Yemen

A more extreme example of a water balance in an arid climate was provided by an investigation in Yemen (Farquharson *et al*., 1996). The topography of Yemen (Fig. 8.16) is dominated by mountain ranges running parallel to the Red Sea coast and rising to over 3700 m. The rainfall is affected by the topography and varies from about 800 mm to the west of the main mountain range to less than 50 mm along the Red Sea and Gulf of Aden coastal plains, and also inland towards the *Rub el Khali* or “Empty Quarter”. The rainfall is confined to two seasons, around March–May and July–September, and the relative importance of these two seasons varies with proximity to the two coasts. Analysis revealed that the annual total rainfall varied with the number of storms rather than their intensity, and storm magnitudes were reasonably uniform over the region. With annual potential transpiration over the area at about 1800–2000 mm, the runoff is extremely intermittent and depends on the incidence of storms. An unusual feature of the land use (Plate 8.4) is that much of the area is subject to bunding and terracing, even on steep hillsides, to detain runoff and permit rainfed agriculture.

Long-term measurements of runoff were rare, but a combination of rainfall modelling, based on the typical number of storms, and runoff modelling based on classification of land types and antecedent conditions, was used to extend several flow records. The seasonal distribution of runoff from the longer records reflected the variation over the country of seasonal rainfall. The inclusion of shorter records in a comparison of mean annual rainfall and runoff depth (Fig. 8.17) showed a fair degree of scatter but no differentiation between different regions. The comparison suggested that the runoff coefficient does not drop below 5% even for the most arid basins. This

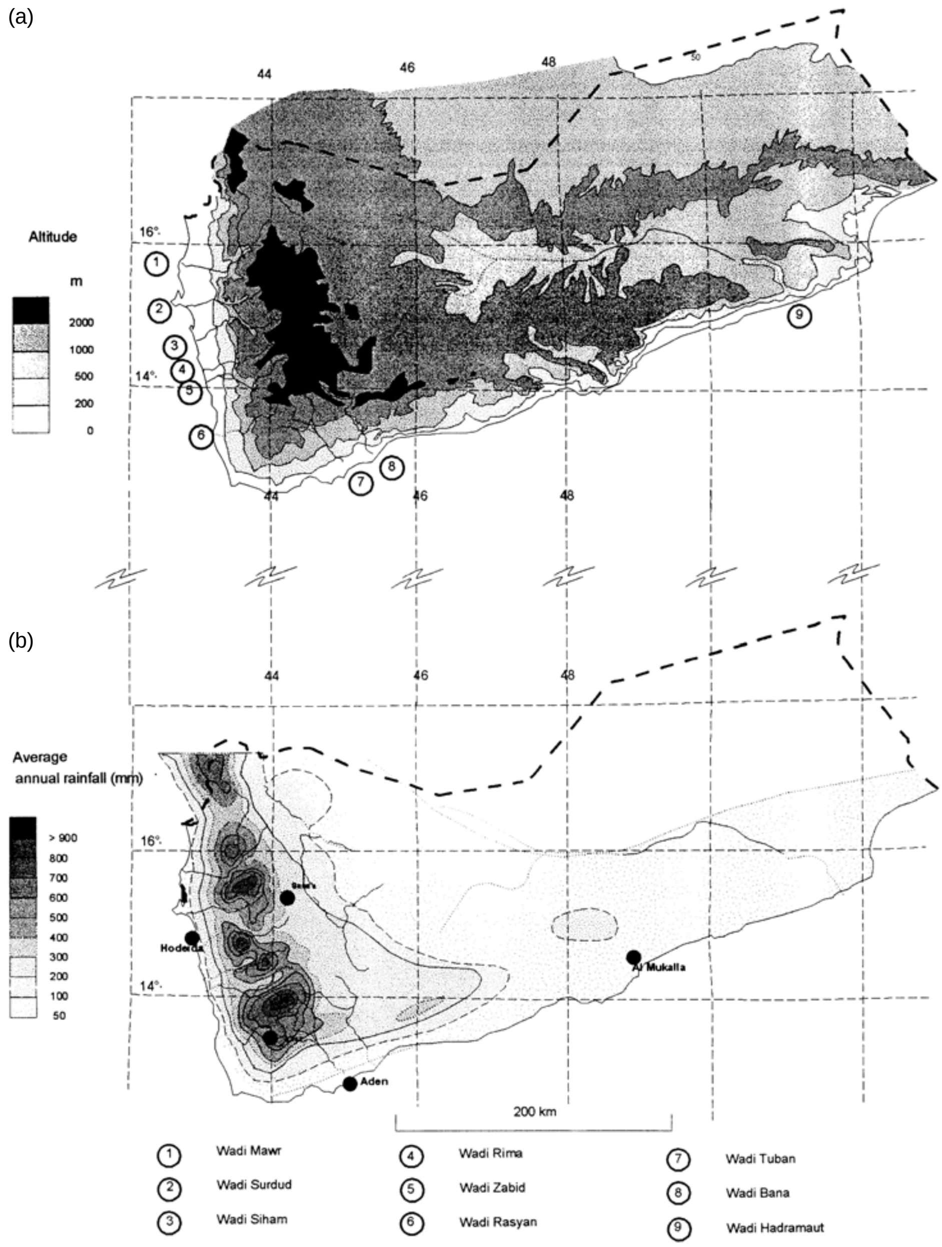

Fig. 8.16 *Yemen: (a) topography and (b) rainfall (from Farquharson et al., 1996).*

does not accord with the expectation that the runoff coefficient tends to zero as the rainfall decreases, but may be consistent with the rainfall model. If the storm magnitudes are not related to annual average rainfall, and antecedent rainfall is low, the runoff generated by each storm will depend on basin characteristics such as geology, which are constant.

***Plate 8.4** Terracing west of Sana'a, Yemen.*

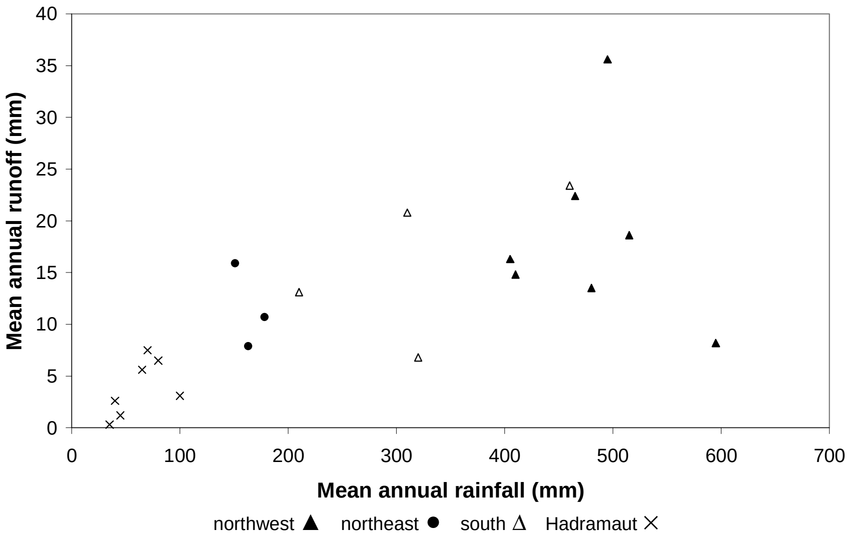

***Fig. 8.17** Yemen: rainfall and runoff (after Farquharson et al., 1996).*

MONSOON CLIMATE

Betwa basin

The water balance of the Betwa basin in central India has been discussed in Chapter 5. In this basin a relatively short rainfall period is followed by a long dry season when the only available moisture is provided by soil moisture storage. Part of the net rainfall during the monsoon season is retained as soil moisture and maintains transpiration during the dry season until the storage is exhausted. The way in which the depth of the soil moisture store can be deduced from the balance between a 50-year measured series of seasonal net rainfall and runoff depth was described in Chapter 5, and a direct method of measuring groundwater recharge was described in Chapter 6. Further studies (Sutcliffe & Green, 1986) were carried out on a tributary of the main river. Within the Nion basin in the centre of the Betwa basin to the east of Vidisha (see Fig. 3.8), detailed measurements of the different components of the water balance were taken over two monsoon seasons. Fortuitously, these two seasons were very different; one was slightly above average in terms of rainfall, while the second was an exceptionally dry year. This investigation was followed by the formulation of a simple hydrological model which expressed the understanding gained of the rainfall–runoff process.

A gauging station was established in 1977 on the Nion about 10 km above the confluence with the Betwa, with a contributing area of 921 km^2; rating was by current meter. Additional rainfall stations were installed to provide six daily records, and basin averages were calculated by the isopercentile method. Daily potential transpiration estimates were derived from meteorological records at Bhopal using a variant of the Penman method. A network of eight soil moisture sites (Hodnett & Bell, 1986) was monitored at frequent intervals from September 1977 to September 1978 as part of the wider programme, with five in wheat, two in pulses and one in rough grass, in proportion to actual land use. In 1978–1979, when rainfall was exceptionally low, a weighted mean was deduced from just four sites. The basin storage was estimated from the mean storage down to a depth of 2.1 m. During the 1978 monsoon, groundwater levels were measured at five observation wells in the basin, and in 1979 were derived by correlation from the one well still operating; the changes in well level were converted to storage change using a storage coefficient estimated from pumping tests to average 0.1. These estimates of basin rainfall, transpiration, runoff, soil moisture storage and groundwater storage are summarised as mm over the basin in Fig. 8.18(a) and (b) for the monsoon months in the years 1978 and 1979. The timing of the rainfall input and its dispersion through storage and runoff are also illustrated clearly by the cumulative analysis of Fig. 8.19. Because the rainfall in the 1978 monsoon, at 1235 mm, was above average, while the 1979 rainfall, at 460 mm, was the lowest on record, the contrast between the two years throws light on the water balance processes. It is clear from the cumulative depths in the different stores that rainfall initially recharges the soil moisture store, and not until this is full does any net rainfall reach groundwater or surface runoff. Once the groundwater store is full, when the water level in this area reaches the surface, all the net rainfall becomes runoff.

In the first phase of the rainfall season of 1978 (days 1–25), rainfall goes almost entirely to soil moisture recharge. In the second phase (26–40), soil moisture recharge continues but the shallow aquifer also responds to groundwater recharge and some runoff occurs. During the third phase (41–98) there is little further change in soil moisture or shallow aquifer storage, and the bulk of the net rainfall becomes runoff; most of the runoff occurs during this phase. In the fourth phase baseflow runoff is maintained by groundwater storage, and actual evaporation from the basin depends on soil moisture and the occasional shower. In fact the recession curve of baseflow can give information directly about groundwater storage; this is particularly useful in a highly seasonal environment, as the volume of runoff after the end of the rainfall season can easily be estimated from the hydrograph and converted to storage depth over the basin.

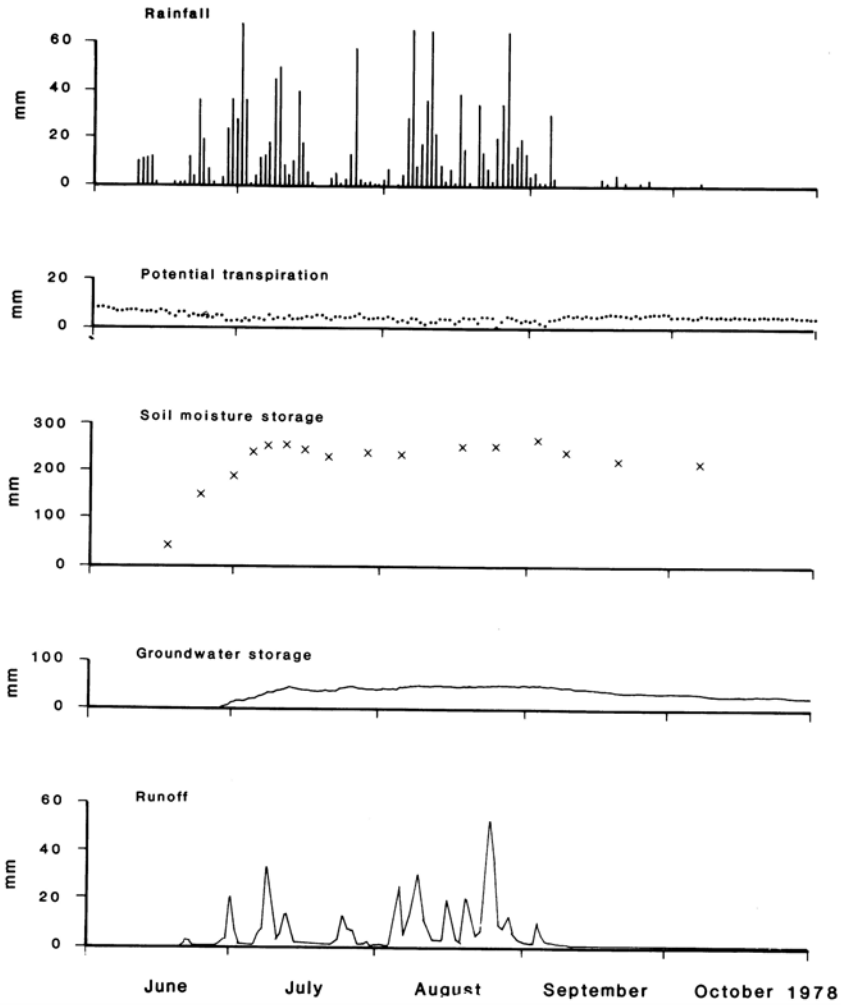

Fig. 8.18(a) *Nion basin, central India: water balance in 1978 (from Sutcliffe & Green, 1986).*

The water balance for 1979 was very different because of the very low monsoon rainfall. However, the sequence of events was similar, and it is the later processes in this sequence, like groundwater recharge and runoff, which were most affected by the low rainfall. Figure 8.19 shows that soil moisture recharge was little lower than in 1978, while the transpiration during the monsoon was below the potential rate. The groundwater recharge occurred later in the season, but was similar in total to that in 1978. Although the runoff depth was severely reduced, the depth of rainfall which occurred before runoff was generated was similar to the previous very different year.

Conceptual model

A simple conceptual model, designed to reproduce the links between the processes, is illustrated in Fig. 8.20. This model was developed from a basic understanding of the physical processes which were dominant within the Nion basin; because this basin was

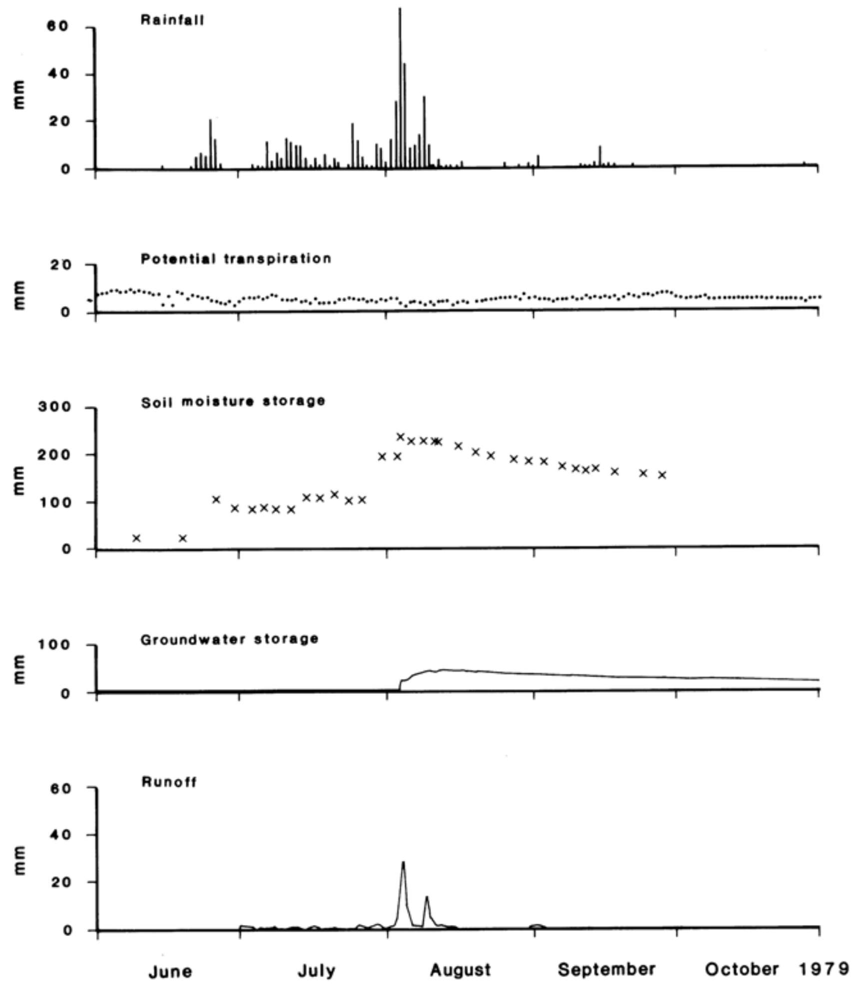

Fig. 8.18(b) *Nion basin, central India: water balance in 1979 (from Sutcliffe & Green, 1986).*

remarkably homogeneous in terms of soils and geology, it was possible to represent it by a lumped model, in which the basin is represented by a single point. The "black cotton soils" overlying weathered basalt provided an uniform soil moisture store which largely controlled the partition of rainfall surplus first into soil moisture storage, then into groundwater and surface runoff. The model was able to represent these successive processes as single storages, which were filled in turn. The model (Fig. 8.20) consisted of a single soil moisture store, a groundwater store, and a channel store. The rainfall, R, is added to the soil moisture store, S; when this exceeds a first threshold, T_1, evaporation occurs at a rate a_1E, proportional to potential transpiration, and drainage, $a_2(S - T_1)$, to groundwater and runoff is proportional to this excess storage. When the soil moisture, S, exceeds a second threshold, T_2, then all the excess drains to groundwater, and to runoff in a proportion fixed by a parameter, a_4, between 0 and 1. When there is water, G, available in the groundwater store, drainage to runoff, a_3, is proportional to this store; when G exceeds a threshold, T_3, the excess runs off directly.

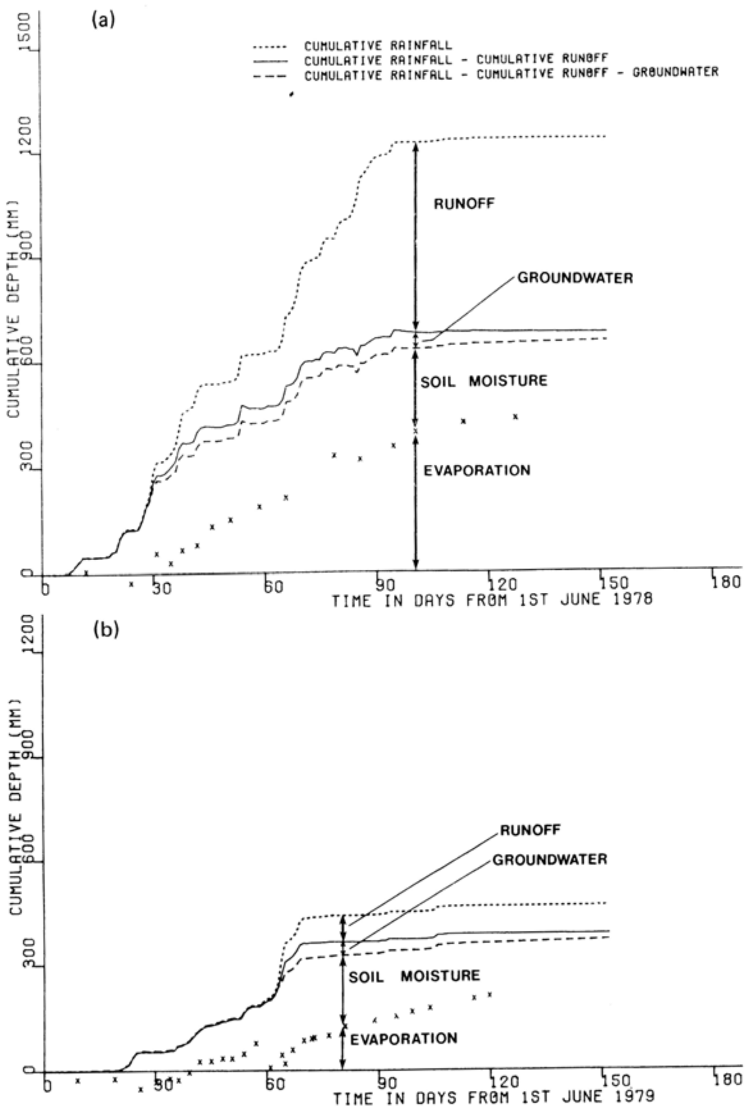

Fig. 8.19 *Nion basin, central India: cumulative water balance in 1978 and 1979: (from Sutcliffe & Green, 1986).*

The runoff is routed through a linear reservoir so that $Q = a_5V$. There was also a small allowance for losses before the monsoon, and the evaporation was reduced during the fallow period at the end of the monsoon.

Table 8.1 *Estimated values of model parameters.*

a_1	1.0	T_1	150 mm
a_2	0.0025	T_2	250 mm
a_3	0.014	T_3	50 mm
a_4	0.65		
a_5	1.8		

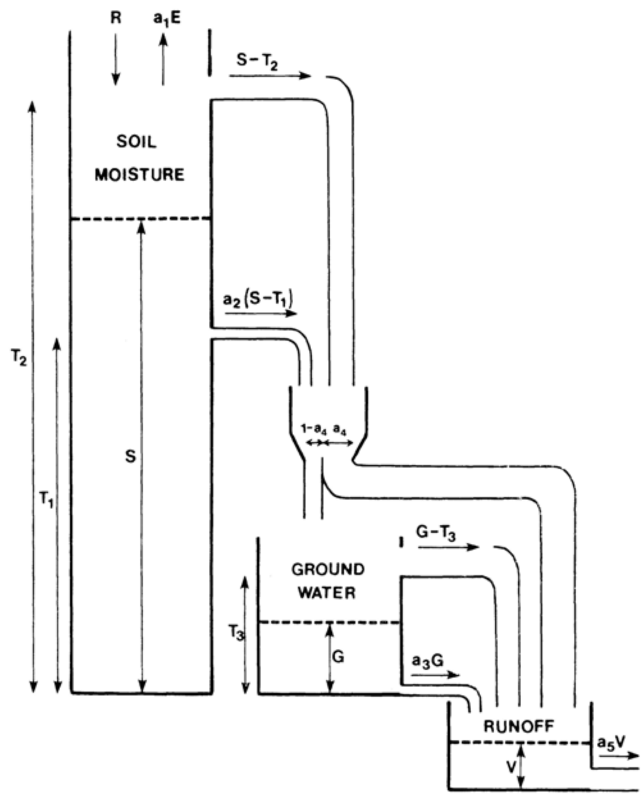

Fig. 8.20 *Nion basin, central India: conceptual model (from Sutcliffe & Green, 1986).*

Although this model has eight parameters, a_1–a_5 and T_1–T_3, it is simple and the parameters have physical meanings. Because direct measurements of soil moisture and groundwater storage were available, some parameters (Table 8.1) were estimated directly and others were adjusted by comparing predictions with measurements. The predicted and measured values of soil moisture storage, groundwater storage and runoff, were found to correspond well in two successive seasons with strongly contrasting rainfall inputs. Comparisons between predicted and observed values in 1978 (Fig. 8.21) suggest that a reasonable representation of the physical processes was achieved. The model was very simple in concept, but was also used elsewhere as a forecasting tool.

Ancient water resources development

The relative simplicity of the water balance in the monsoon climate of the Betwa basin, and the evident benefits of ensuring the availability of water throughout the long dry season, doubtless led to the construction of a series of reservoirs which have been dated to the period starting about 150 BC in this area. During an archaeological survey of an area within about a 15 km radius of Sanchi hill, near Vidisha, a centre of Buddhist propagation in central India, a set of 16 ancient dams was discovered (Shaw, 2000); these were almost all dated to between the 2nd–1st centuries BC and the 5th century AD, using *naga* serpent sculptures associated with the structures, stone facing and associated archaeological sites. Simple estimates of the reservoir volumes associated

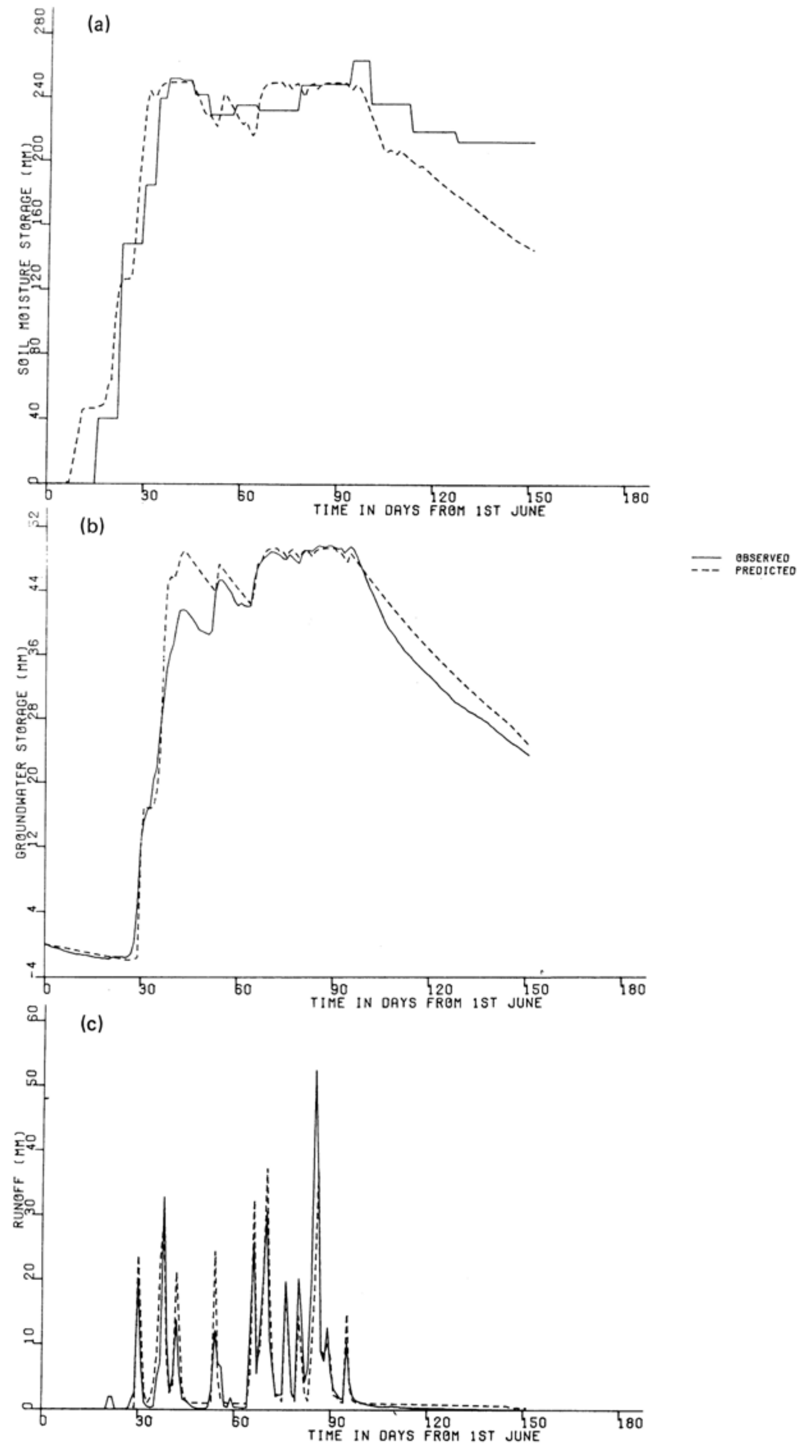

Fig. 8.21 *Measured (—) and simulated (- -) soil moisture storage (a), groundwater (b), and runoff (c) for the Nion basin (from Sutcliffe & Green, 1986).*

Plate 8.5 *Masonry on the upstream side of the Morel Kala dam, central India.*

Plate 8.6 *Spillway on the Devrajpur dam, central India (from Shaw & Sutcliffe, 2001).*

with these dams were made from measurements of dam heights, lengths and reservoir gradients. When these were compared with runoff volumes derived from basin areas and runoff depths from the Betwa study, it was found that the reservoir volumes were closely related to the runoff volumes (Fig. 8.22), which suggested that the dams were designed with an understanding of the principles of water balance. It is suggested (Shaw & Sutcliffe, 2001) that the reservoirs were designed to irrigate rice for the increased population related to the Buddhist expansion in the Sanchi area. Most of the dams are protected on the upstream face with well-constructed masonry (Plate 8.5) and at least two of the later dams are also protected by spillways (Plate 8.6) which appear to have been designed to pass the 50-year flood (Shaw & Sutcliffe, 2003).

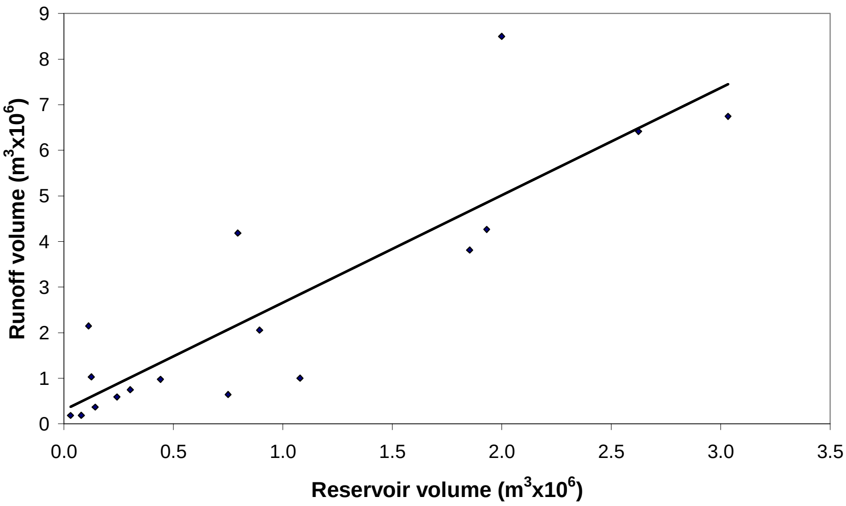

Fig. 8.22 *Ancient dams near Sanchi: reservoir volume versus runoff volume (after Shaw & Sutcliffe, 2001).*

West Africa

The methodological findings of the Betwa study were applied to a regional investigation of water resources in Guinea and Togo-Benin in West Africa (Sutcliffe & Piper, 1986), where the seasonal rainfall distributions were similar to Betwa and the deciduous woodland suggested that the soil moisture regimes were similar. In reviews of potential hydroelectric power resources, runoff estimates were needed at a number of potential reservoir sites over the region.

From monthly series of rainfall and potential transpiration, seasonal net rainfall series for basins were derived and compared with annual runoff totals at a number of sites. These were found to correspond reasonably well in the case of the Guinea stations (Fig. 5.6, for example), generating curves with a slope of about 45° and an intercept of about 200–300 mm. This corresponded well with the similar situation in the Betwa basin. In the case of the stations in Togo and Benin, the correlation between net rainfall and runoff was reasonable, but the slope was less than 45^o. This response was attributed to the fact that much of the area has two rainfall seasons divided by a drier period, so that the soil moisture recharge is not constant from year to year.

Isohyetal maps of gross and net rainfall were drawn (Fig. 5.7(a) and (b)), and mean annual net rainfall was compared with mean basin runoff depth for those sites with reasonably long records (Fig. 8.23). These showed that mean runoff corresponded with mean net rainfall, less 300 mm soil moisture recharge, in Guinea. Annual runoff series could be estimated from seasonal net rainfall series with an allowance of 300 mm for annual soil moisture recharge.

In a later study of the water balance of Sierra Leone, where the climate and the seasonal distribution of rainfall were similar to Guinea, it was found possible to make reasonable estimates of average annual runoff by comparing average rainfall given by an isohyetal map with estimates of potential transpiration. The runoff estimates prepared using the methods developed for Guinea were consistent with the sparse measurements of river flow in Sierra Leone.

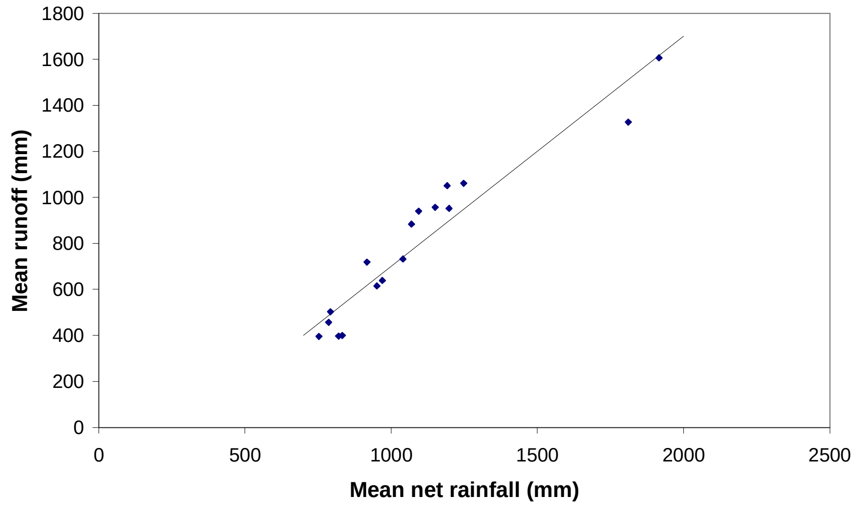

Fig. 8.23 *Guinea: mean annual net rainfall and runoff (after Sutcliffe & Piper, 1986).*

Summary

The simplifying assumption of a constant annual soil moisture recharge can only be applied where the monsoon rainfall can be treated as a single season without a significant period of drought; this is in accordance with the physics of the water balance. These examples of basins in different climates have shown that water balance studies are comparatively simple and most useful in humid areas and in areas of single-season monsoon climates, but are more difficult to apply in semiarid areas.

SOME PRINCIPLES OF HYDROLOGICAL MODELLING

Some examples of hydrological models have been introduced as part of the discussion of water balance assessment. These examples have all tended to simplify the processes involved, and it can be argued that this is a great advantage when dealing with developing countries where data are usually confined to rainfall records, meteorological data from which potential transpiration or evaporation rates may be estimated, and flow records which may not always be numerous or of long duration. Some advantages of simple models and problems of parameter estimation are discussed in this section.

Simple conceptual models

In an early example, Nash & Sutcliffe (1969) argued for simple conceptual methods in forecasting as a compromise between the assumption that streamflow could be separated into baseflow and storm runoff, deriving from effective rainfall, and the argument that it is sufficient to apply physical laws to the physical data of a given catchment in order to forecast discharge. It was noted that river flow could be forecast from rainfall and the present state of the basin, but that practical needs favoured a model based on previous history, in which the present state is itself forecast from the history of rainfall, potential evaporation and observed or computed discharge.

It was argued that if a model is to help understanding of the working of the hydrological cycle and the relative importance of the different elements, it is essential to obtain a guide to the relative significance of model parts and parametric values. It is desirable that a model should have as few parameters as possible, and in particular that two or more parts with similar effects should not occur in the same model; on the other hand, each additional part of a model should substantially extend the range of application of the model. If certain functions of the model could be isolated and its parameters optimised separately, a higher accuracy of determination should be obtained.

Ray at Grendon Underwood

This approach was applied to the records of the Ray at Grendon Underwood, England, where an experimental basin had been established in 1962 on a small area (19 km^2) of deep clay in the upper Thames to allow the development of forecasting techniques.

Occurrences of substantial rainfall were noted over a four-year period, and were related to the runoff volumes estimated between successive recession curves. A simple regression of runoff in 221 periods on the corresponding rainfall was obtained to judge the adequacy of subsequent models; the slope of the regression was found to be 0.31 with a standard error of 0.026.

The first model to be tested assumed a vertical stack of horizontal soil layers, each containing 25.4 mm of moisture at field capacity, with a total storage Z. Evaporation from the top layer occurs at the potential rate, then from the second layer at the potential rate multiplied by C, and so on. When rainfall exceeds evaporation, a fraction H of the excess contributes to generated runoff, and of the remainder anything in excess of a threshold value Y also contributes to generated runoff; the remainder percolates down to soil moisture storage. The potential evaporation is estimated from the Penman formula and multiplied by a parameter T to allow for systematic error. Thus the model has five parameters, some of which could be made inoperative.

After further discussion of the fitting of a model and a proposal for the definition of the efficiency of a model (Nash & Sutcliffe, 1970), this model was then compared (Mandeville *et al.*, 1970) with a second soil moisture model based on a study by Penman (1950) of the Stour; this study has been described in Chapter 6. This model divided the basin into three areas of grass, woodland and a riparian zone, and assumed that in each zone vegetation transpired at the potential rate until the wilting point was reached throughout its rooting depth and thereafter transpired at one tenth of the potential rate. The freely available soil moisture depended on the depth of rooting and soil characteristics or "root constant", and was allowed to vary between the areas of grass remote from streams, areas of woodland with deeper rooting range, and the riparian zone where available water was assumed unlimited. The total area of the basin (18.56 km^2) and the area of woodland (3.44 km^2) were known. The area of grass remote from the river (A) and its root constant (D) provided two parameters. The root constant of the woodland was found to be irrelevant in this climate. As in the other model, the ratio (T) of potential evaporation to open water evaporation was usually fixed. The elements corresponding to H and Y were at times used in this model also. The basis of this model is illustrated in Fig. 8.24.

It had been hoped, perhaps naively, that comparisons of these two soil moisture models would throw some light on which model more nearly represented reality. However, they were in fact similar in their operation. Both allowed the ratio of actual to potential evaporation to fall as the soil moisture deficit increases. However, the

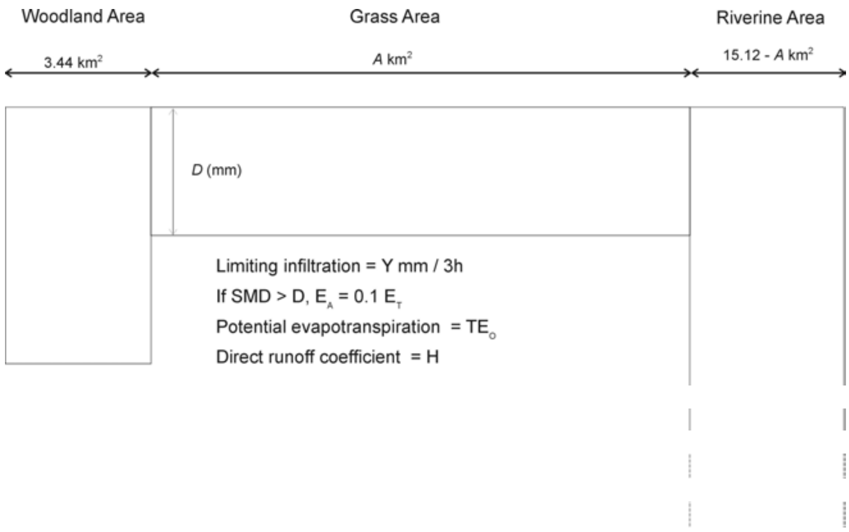

Fig. 8.24 *Basis of the conceptual model.*

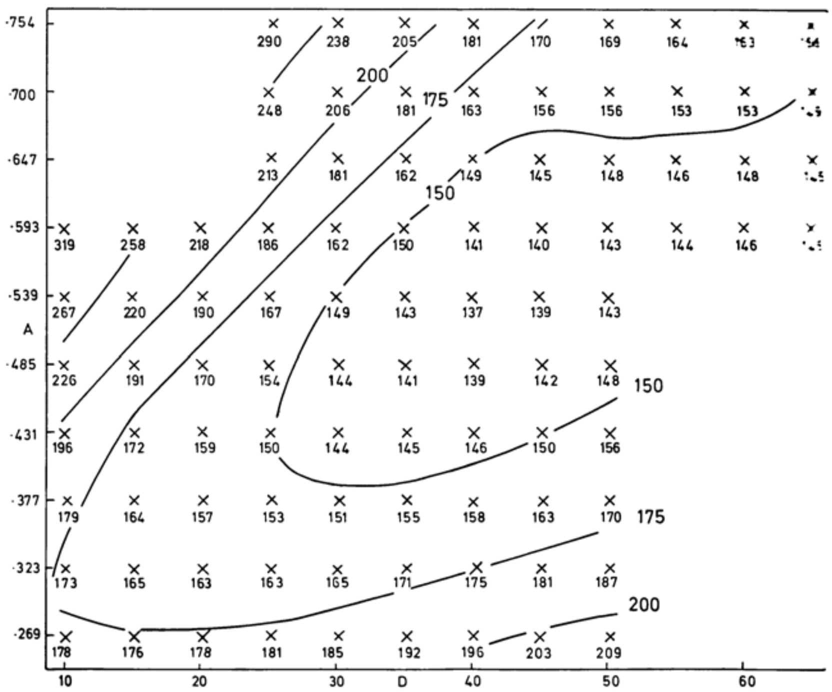

Fig. 8.25 *Dependence of the error function on A and D (from Mandeville et al., 1970).*

second model seemed to be more flexible as it allowed evaporation almost to cease from the grass area while the deficit continued to build up elsewhere.

The regression of the runoff amounts on the corresponding rainfall depths accounted for 59% of the original variance. The same data were used in a trial of the standard forecasting tool of the period, a coaxial graphical correlation relating runoff to rainfall depth, calendar month and antecedent precipitation index, which explained 81% of the initial variance. A series of trials were carried out with the two models, and for each trial the starting and optimised values for each parameter, the optimised variance, the efficiency and the total computed runoff were listed. It was found that with the second model, with one parameter fixed, and with three optimised, plus the initial deficit, the efficiency reached over 90%, which was a distinct improvement over earlier techniques.

However, comparisons of the error functions corresponding with different values of two parameters, with the other parameters fixed at reasonable values or made inoperative, showed that there was clear interdependence between pairs of parameters. In some cases the results of the optimisation, which was basically that developed by Rosenbrock (1960), were affected by starting values and local minima were reached. Figure 8.25 shows the error function surface for the second model as a function of *A*, the grass area, and *D*, the root constant. There is a valley with a flat floor, with no well-defined optimum, running diagonally across the plane. In such cases, it is possible to fix one parameter at a reasonable value after inspection of the basin, though it may be misleading to assume that the parameters correspond to their physical counterparts.

The conclusions of this comparative study were that it is possible to predict river flows with reasonable precision from rainfall and evaporation records provided that river flow data can be used to calibrate a simple model; that the calibration of a model is made easier if the model is designed to exclude parameters with similar effects on forecast flows; that proliferation of parameters requires additional measurements if their calibration is to be realistic; that it may not be possible to choose between different interpretations of hydrological processes by comparisons of models alone.

WATER BALANCE OF LAKES

Studies of the water balances of lakes have advantages and problems of their own. Because lake levels are often available for periods before other hydrological records were started, the study of lake water balances can extend the period of information about the water resources of a region. Where lakes form the limit of an internal drainage system, lake levels can give direct indicators of variations of water resources before the availability of inflow and outflow records. This has been true of Lake Turkana, which receives inflow from the Omo valley in Ethiopia and from the Turkana River in northern Kenya. Historic levels have been inferred from investigations by Butzer (1971) who was able to use the maps compiled by early travellers. These levels support similar trends in other parts of East Africa. The former levels of other lakes in the Ethiopian Rift Valley have been investigated by Grove *et al.* (1975) and given insight to climatic change over this key area. Even where lakes have outlets which have only been recorded for limited periods, they can provide a useful check on regional hydrological records. During a study in 1986 of the water resources of rivers flowing into Lake Abaya in the Rift Valley, it was possible to use lake levels to confirm the validity of flow records and also the derivation of rainfall–runoff relations used to infer the flows of minor tributaries. Nearly complete records of inflow to the lake were compared with lake rainfall and evaporation to provide predicted lake levels over a period of nine years; when these were compared with measured lake levels a period of apparent inconsistency was noted, but this was found to coincide with a rise in lake level which resulted in a period of high flows recorded in a river flowing from Lake

Abaya to Lake Chania downstream. Thus the downstream lake evidence supported the validity of the upstream water resources study.

There are channels leading from these two lakes towards Lake Stefanie, which is normally an area of grass and wetland; there is evidence that on rare occasions Lake Stefanie also becomes a true lake, and these periods coincide with high levels of Lake Turkana and Lake Victoria, which are known to have occurred around 1878 and 1961.

Lake Victoria

The water balance of Lake Victoria has been the subject of a number of studies because of its importance as the source of the White Nile, and because of the dramatic rise in lake level and outflow which occurred after 1961–1964.

Lake levels have been measured regularly since 1896. From 1896 to 1960, the levels fluctuated between 10.2 and 12.0 m on the Jinja gauge (Fig. 8.26), but rose 2.5 m to over 13.0 m in 1961–1964. Since 1964 the general trend has been downwards, but there were rises in 1979 and 1998. The lake outflows were controlled naturally by the geometry of the Ripon Falls before 1951, and the levels and outflows were linked by a stable control. Since 1951, when the construction of the Owen Falls dam began some 3 km downstream of the Ripon Falls, the outflow has been controlled to match the earlier relation between lake level and outflow, so that the dam is operated on a run-of-river basis. This relation was deduced from current meter gauging below the lake, but was extended by hydraulic modelling of site conditions to match higher levels. A composite outflow record has been deduced (Sene & Plinston, 1994) and is brought up to date in Fig. 8.27.

Some 20 tributaries into Lake Victoria drain a total area of 190 000 km^2 compared with a lake area of 67 000 km^2. The Kagera, draining much of Rwanda and Burundi, has been measured since 1940. Several Kenyan tributaries have been measured since 1956; nearly all of the other tributaries were measured from 1969 to 1978. The total tributary inflow for 1956–1978 was estimated from these records. The inflow (Fig. 8.28)

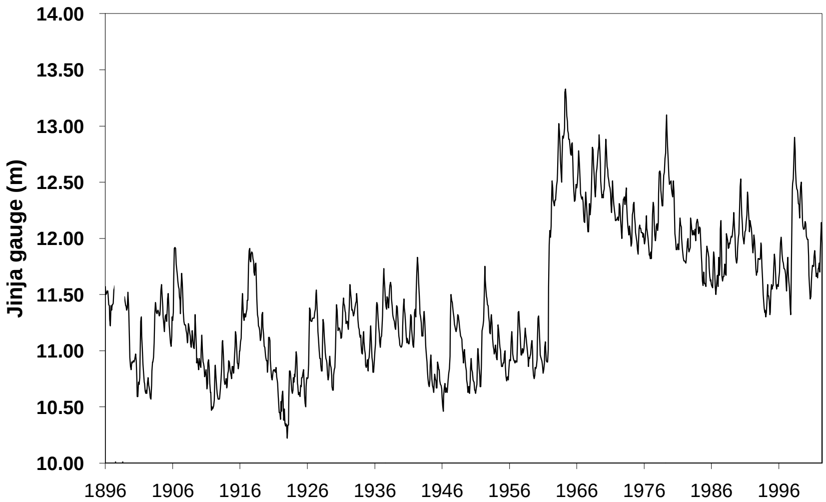

Fig. 8.26 *Lake Victoria lake levels: Jinja gauge, 1896–2000.*

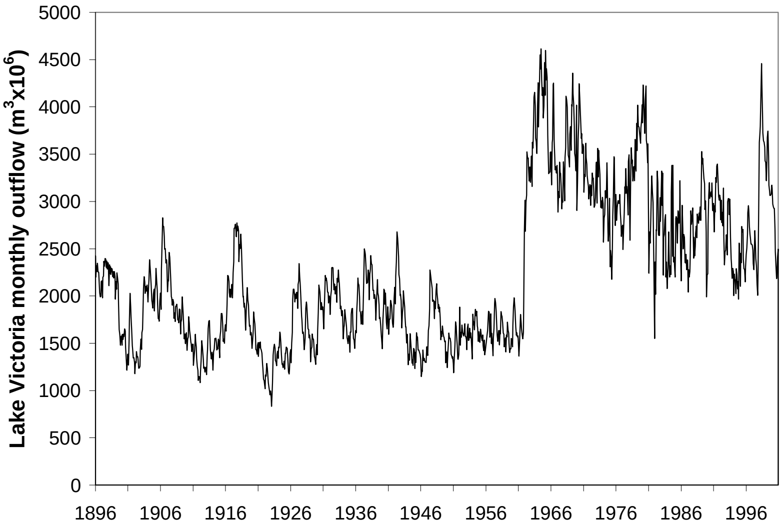

Fig. 8.27 *Lake Victoria outflow: 1896–2000.*

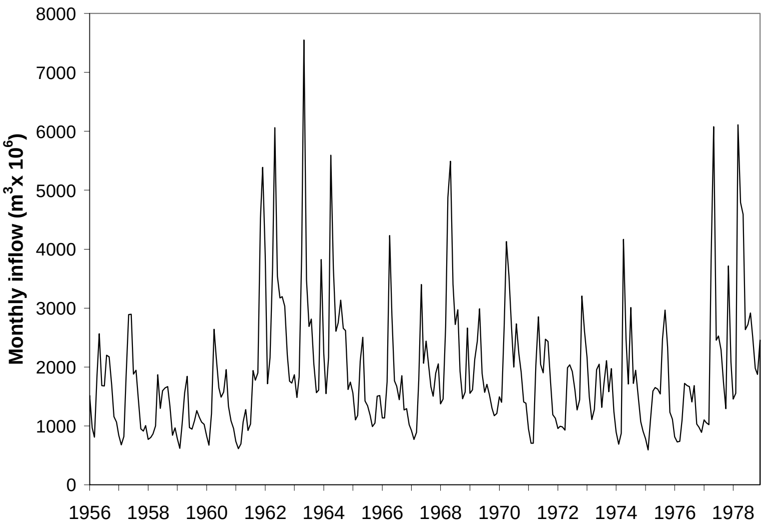

Fig. 8.28 *Lake Victoria tributary inflow: 1956–1978.*

increased markedly after the heavy inflows of 1961–1964. This inflow was variable from year to year as the natural variability of catchment rainfall was supplemented by the sensitivity of the runoff coefficient to rainfall.

The main component of the input to the lake is in fact the rainfall over the lake surface; indeed this rainfall has been described as the true source of the Nile. The interaction of lake breezes with the prevailing easterlies results in higher rainfall over the lake than on the surrounding basin, with rainfall in the morning on the eastern shore and evening rainfall on the west. This has been confirmed by some measurements on islands in the lake. It is not easy to develop a long-term series of lake rainfall, as the only continuous stations are at eight sites around the lake.

However, a reasonable series was deduced (Piper *et al*., 1986) by using these stations as indices of monthly and annual rainfall, while enhancing the variability to counteract the effect of averaging the indices, and multiplying the series by the mean lake rainfall derived from the water balance. There is evidence from the rainfall records around the lake that there was a significant rise of rainfall after 1961–1964 in the October–November season, which has evidently been sufficient to affect the lake balance.

As it is difficult to distinguish variations in rainfall from those of lake evaporation, the lake evaporation was taken as constant from year to year, and was based on monthly average open water evaporation derived by the Penman method for stations around the lake.

The annual means of all these water balance components for the period 1956–1978 are expressed in Table 8.2 as mm over the lake area, and their variations are expressed in terms of standard deviation and coefficient of variation. Table 8.2 confirms that rainfall is the major component of the balance, but also that tributary inflow, though only about 15% of the total input, has a much higher relative variability and is therefore an important factor in the rise of the lake. The lake balance is extremely sensitive to increases of lake rainfall and related tributary inflow, as the average rainfall is of the same order as lake evaporation; an increase in input has a disproportionate effect on the outflow, which has a variability somewhat higher than the inflow alone.

Table 8.2 *Annual water balance of Lake Victoria, 1956–1978 (after Sutcliffe & Parks, 1999).*

Year	Lake rainfall (mm)	Tributary inflow (mm over lake)	Lake outflow (mm over lake)
Mean	1858	343	524
SD	180	103	149
CV(%)	9.7	30.1	28.4

Table 8.3 *Monthly components of Lake Victoria water balance, 1956–1978 (after Sutcliffe & Parks, 1999).*

Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Year
Lake rainfall (mm)												
138	153	228	309	208	71	52	68	104	157	201	170	1858
Total inflow (mm over lake)												
20	16	23	41	48	31	30	31	29	23	25	26	343
Lake evaporation (mm)												
135	135	145	130	125	120	125	135	140	145	130	130	1595
Lake outflow (mm over lake)												
43	39	43	44	47	47	46	44	43	43	41	44	524
Lake level (m on Jinja gauge)												
11.93	11.93	11.99	12.15	12.23	12.16	12.06	11.98	11.92	11.90	11.97	12.01	

The seasonal lake balance is summarised by Table 8.3, where the components of the balance and the resulting lake level and outflow are given. The annual balance is illustrated by Fig. 8.29, where the observed and predicted annual lake levels are compared. It will be seen that the balance is not only close over the whole period, which is inevitable from the method of deriving rainfall, but also follows the pattern of rises and falls between these dates. The result of the change in the lake balance has been the doubling of the lake outflows after 1961–1964. This rise has been reinforced by the net effect of the basins of Lake Kyoga and Lake Albert (Sutcliffe & Parks, 1999), and has therefore had a dramatic effect on the wetlands of the Sudd downstream.

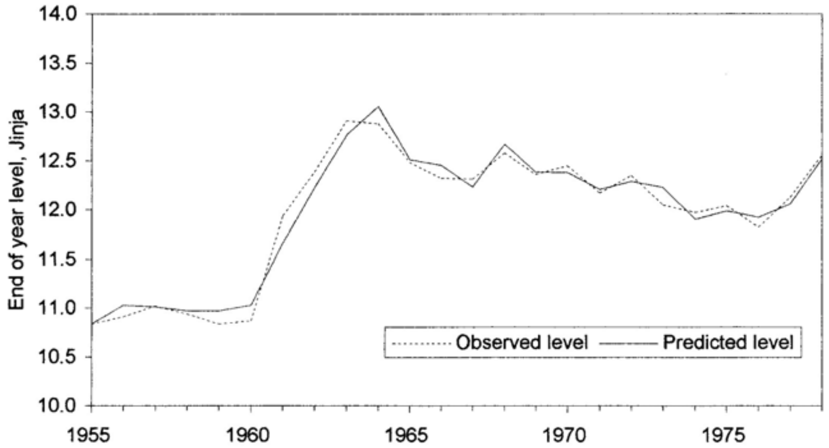

Fig. 8.29 *Lake Victoria annual balance, 1956–1978 (from Sutcliffe & Parks, 1999).*

It has been possible to use indirect evidence to take the study of Lake Victoria hydrology back before the start of historic lake levels. Lyons (1906) had visited the East African lakes and gathered oral and published evidence at that time about earlier lake levels. There were a number of travellers' observations of exceptionally high lake level in August 1878, a fall of about 2.5 m by 1891, followed by a rise in 1892–1895 and a period of falling levels from the start of observed levels in 1896. The 1878 peak was supported by observations of flooding at Lado below Lake Albert, where levels were still high in December 1878.

These high levels were supported by evidence from downstream. There is evidence of widespread flooding in the Sudd in 1878 and 1895, while river levels at Khartoum in May 1879 indicated high lake outflows in 1878. Observed flows at Aswan were high on several occasions in the period 1870–1898, and it has been possible to reconstruct early levels from downstream evidence (Tate *et al*., 2001). There is archaeological evidence for an upper bound to lake levels (Bishop, 1969). The 1964 peak levels were within 60 cm of the level of gravel deposits with dated charcoal fragments in Hippo Bay cave. Bishop (1969) concluded that "at no time since 3720 ± years ago has Lake Victoria risen to re-occupy the 10- to 12-foot cliff notch" (a level equivalent to 13.85 to 14.46 m on the Jinja gauge).

It is also possible to link modelling of Lake Victoria to climate changes in rainfall and evaporation to estimate potential impacts on lake water balances and levels (Tate *et al*., 2004).

WETLAND WATER BALANCE

Wetlands play an important role in a number of major river systems. They have received considerable attention in the Nile basin, where the loss to evaporation of about half the inflow of the Bahr el Jebel before it reaches the White Nile has drawn attention to the hydrology of the Sudd. Proposals to divert the part of the inflow to the Sudd around the swamps in the Jonglei Canal have been studied in various forms since 1904, and the benefits and the impacts of this project are considered in Chapter 11. Here, the problems of carrying out water balance studies in major wetlands are discussed and illustrated by studies of the Sudd and other major African wetlands.

The major difference between water balance studies of wetlands and lakes are that the transpiration from swamp vegetation has been less studied than open water evaporation, and therefore there is less certainty about the rate of transpiration or its relation to open water evaporation. Methods are now available to measure evaporation rates above different surfaces, and these should in the future provide means of refining wetland water balances; however, up to the present it has been necessary to apply the general principles of water balance. It is also true that while the relation between lake area and volume of water stored can be deduced from surveys around the lake, together with depth surveys within the lake, it is not a simple matter to derive relations between wetland area and water volume, even with sophisticated methods of survey. The problem is similar to deriving area–volume relations for a reservoir, with the added problems that access is not easy, that water surfaces can vary in level over an area, and that both water surfaces and water depth are usually obscured by vegetation. Indeed, the water depth in a papyrus swamp is not well defined, and even when using a probe must be judged by a change of resistance. It is likely that a comparison of swamp area, revealed by satellite imagery, with volumes stored derived from water balance studies, could now present a means of deducing area–volume relations, but it may not be easy to find periods when satellite availability and precise swamp water balances coincide.

The Sudd

In the case of the Sudd, there is some evidence from surveys that a linear relation between area and storage is reasonable. This is equivalent to an exponential relation between water surface elevation and area, and thus to a constant average depth of water over the flooded area. This is not an unreasonable simplification; it has the incidental advantage that there is in this case a direct solution (Sutcliffe & Parks, 1987, 1989) to the water balance equation linking successive areas and volumes of flooding, whereas other forms of relationship require an iterative solution to allow for evaporation from a changing area.

The equation of continuity for a time interval dt is:

$$\mathrm{d}V = [Q - q + A(R - E)]\mathrm{d}t - r\mathrm{d}A$$

where V is volume of flooding, Q is inflow, q is outflow, R is rainfall, E is evaporation, A is flooded area and r is soil moisture recharge, which is positive but set to zero when dA is negative. Percolation to groundwater is unlikely to be significant in the case of the Sudd. Field evidence suggests that a linear relation between V and A is reasonable for the Sudd, and this can be expressed as $A = kV$ as $A = 0$ when $V = 0$. This implies a constant average flooding depth of $1/k$ m, and a value of $k = 1$ was deduced from personal observation and adopted after testing different values and also different forms of the relationship.

The analysis for month i, allowing for rainfall and evaporation over the mean of the initial and final values of area A, and substituting kV for A, becomes:

$$V_{i+1} = V_i + Q_i - q_i - k/2(V_i + V_{i+1})(E_i - R_i) - kr_i(V_{i+1} - V_i)$$

and hence

$$V_{i+1} = (V_i[1 + k\{r_i - (E_i - R_i)/2\}] + Q_i - q_i)/[1 + k\{r_i + (E_i - R_i)/2]$$

In the case of the Sudd, monthly inflows and outflows were known from 1905 to 1980, and a rainfall series was compiled from nine stations around the swamps. Evaporation

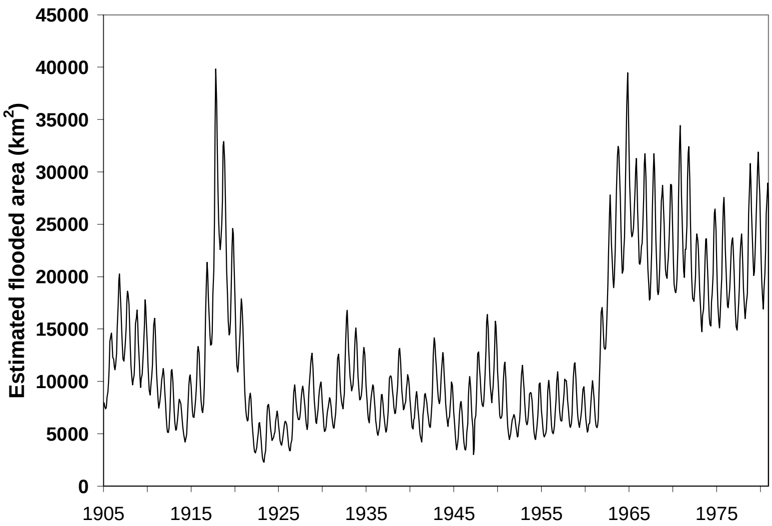

Fig. 8.30 *Bahr el Jebel swamps: monthly estimates of flooded area, 1905–1980 (after Sutcliffe & Parks, 1987).*

was taken as open water evaporation estimated for Bor by the Penman method. Soil moisture recharge was taken as 200 mm at the start of the wet season, and reduced by net rainfall during the wet season.

Storage volumes and flooded areas were estimated at monthly intervals starting from initial estimates of 8000 $m^3 \times 10^6$ and 8000 km^2, respectively. These values corr-esponded reasonably with estimates of flooded areas derived over the years from survey maps, vegetation maps and satellite imagery. The series of monthly flooded areas (Fig. 8.30) showed the influence of the seasonal torrents in providing annual areas of flooding and recession, which provide the important areas of seasonal flooding, while the lake-fed steady flows of the Bahr el Jebel below Lake Albert are responsible for the more permanent swamp. However, the important effect of the rise in Lake Victoria after 1961–1964, when the areas of permanent swamp more than doubled, is clearly evident.

African wetlands

A similar analysis was applied (Sutcliffe & Parks, 1989) to provide a comparison between four major African wetlands: the Sudd of the Bahr el Jebel; the Okavango swamp in Botswana; the Niger interior delta in Mali; and the wetlands of the lower Senegal basin. The main differences between the four wetlands were the seasonal distributions of river inflow. Those of the Sudd were a combination of lake-fed steady flows and highly seasonal torrents. The inflow to the Okavango was seasonal but damped in the long course from the mountains of Angola. The inflows of the Niger and Senegal are both derived from the highly seasonal rainfall in their headwaters in the mountains of the Fouta Djallon in Guinea; they are therefore markedly seasonal with the flows concentrated in 3–4 months.

The inflows and outflows, together with average rainfall and estimates of open water evaporation, were used in the model developed for the Sudd to generate

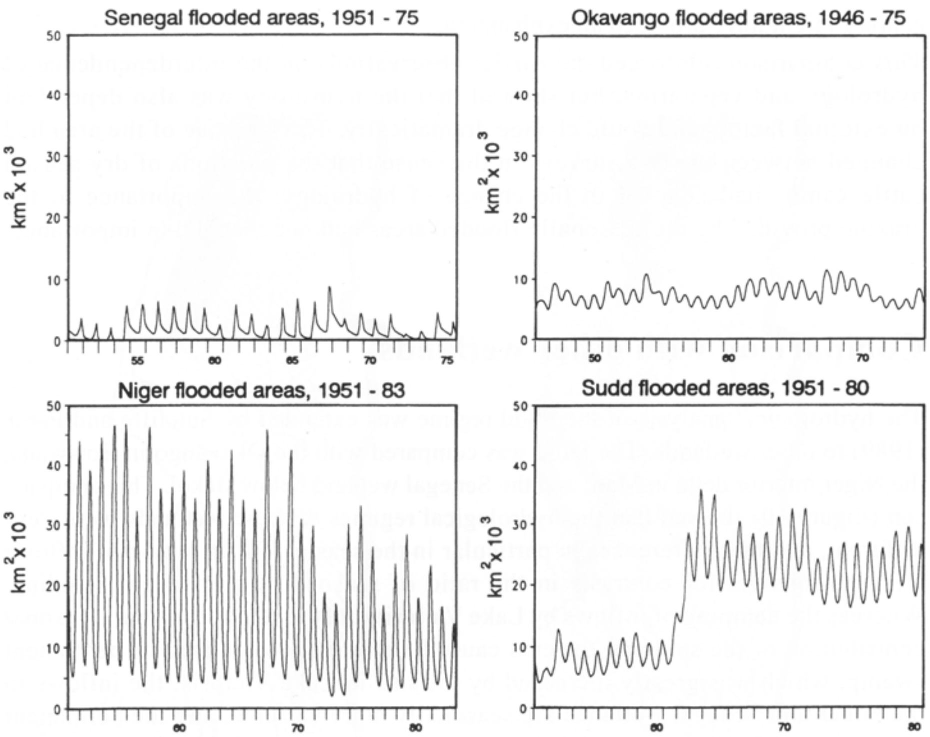

Fig. 8.31 *Comparison of flooded areas (km^2 x 10^3) in Senegal, Okavango, Niger, Sudd (from Sutcliffe & Parks, 1989).*

estimates of flooded areas over similar numbers of years. The only difference was that an average depth of flooding of 2 m rather than 1 m was assumed for the Senegal wetland. The results (Fig. 8.31) showed that what appeared to be areas of similar status on the map had quite different flooding regimes. The Okavango swamp had similarities with the Sudd, in that there was a significant area of permanent swamp, though the effect of Lake Victoria was missing. The wetlands of the Niger and Senegal, however, were much more seasonal with small areas of permanent flooding. In contrast to the increased flooding in the Sudd after the rise in Lake Victoria, the Niger and Senegal wetlands have decreased in extent after about 1970, as a result of the effects of the Sahel drought on river inflows.

The vegetation corresponding to wetlands of different hydrological regimes is discussed in Chapter 11; however, the contrast between the flooding regimes of these major African wetlands is reflected in their vegetation. Those with high proportions of permanent swamp are naturally more dominated by papyrus, while those with a higher proportions of seasonally flooded land are dominated by grasses. These different regimes are controlled by the inflows to the wetlands rather than the local climates. The Sudd and Niger wetlands are similar in terms of latitude and seasonal climate; the key difference is the large lake-fed inflow to the Sudd which feeds the permanent swamp.

CONCLUSION

These examples illustrate the general point that simple water balance models can be very effective in explaining the physical processes operating in a river basin, and can lead to an understanding of the runoff provided that the climate, the physical hydrology and the vegetation are carefully considered. Comparisons of precipitation and potential transpiration can lead to a simple understanding of different hydrological regimes, but the seasonal timing is as important as a comparison of annual totals. Provided that an understanding of the main driving forces is incorporated, simple models of the water balance can be extremely useful in explaining the hydrology of an area, and can provide a useful check on more complex hydrological modelling and on the basic data. In fact in these days of increasing complexity of analysis and increasing reliance on computer manipulation of data, it is extremely important to ensure that the hydrological forces controlling runoff are understood and incorporated in analysis, rather than relying on more complicated sets of parameters and relationships in attempting to explain runoff.