Environmental Hydrology and Hydraulics

Eco-technological Practices for Sustainable Development

S.N. Ghosh
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Preface

The motivation for writing this book emanated from our firm belief that for sustainable development we need to conserve all our natural resources such as air, water etc. In addition, we need to emulate all the natural processes to the maximum possible extent in all our design endeavors aimed at achieving development through an improved infrastructure. Individuals and organizations at all levels should be made aware of the fact that water is a precious natural resource which is crucial to our survival. Water needs to be judiciously used in the context of an increasing population—not only to sustain essential requirements such as those for drinking and domestic usage but also for increased food production, industrial usage, power generation, navigational requirements, pisciculture, recreation, landscaping etc.

In view of the uncertainties associated with the global hydrological cycle over which human beings do not have any control, there is the problem of scarcity as well as excess of precipitation both spatially as well as temporally across the globe. Since precipitation is the primary source of water it is essential to harness/store this precious water resource for various usages at all the times as well as at all possible locations.

In the developed world, a large number of hydraulic structures have already been built to augment the water availability so that their overall water demand consistent with their industrial growth and their standard of living can be met. On the other hand, the developing/under-developed world consisting of low and medium income group of nations and accounting for about 85% of the global population have not been able to develop sufficient built-up capacity to augment/store their naturally available water resources. As a result of this most of their precipitation water flows into seas/oceans during peak flow season.

In view of the huge costs involved in major water resources projects as well as on account of their other associated operational, social and environmental problems, such as long gestation period, rehabilitation of project affected population, submergence of valuable forests, loss of rare plant/animal species and minerals, there is a strong international opinion against large scale water resource development. As an alternative to this, a greater emphasis is being given at all levels to small scale development and efficient use of water resources through appropriate technologies inclusive of revival of traditional technologies, water harvesting, low-cost and biological treatment/reuse of wastewater.

Already there are many books dealing with hydrology, hydraulics and hydraulic structures, which generally deal with larger problems of development, analysis, design and implementation of water resources. But there are not many books which deal with small-scale development of water resources consistent with the environmental concerns as well as application of relevant eco-friendly technologies.

Keeping all these factors into consideration, this book has been grouped into five chapters. Each of these chapters is briefly described here.

Chapter 1 provides the reader with a basic background of hydrology as well as all the hydrological processes. It also tries to establish the linkage between imbalance in these hydrological processes due to human intervention and its impact on ecology.

Chapter 2 deals with various uses of water in the order of importance to human civilization. Many relevant case studies from the developing as well as developed world are elaborated to describe water use for municipal requirement in urban/rural neighborhoods, agriculture, industries, hydropower generation, navigation, fisheries as well as recreation.

Chapter 3 explains the three fundamental conservation principles of hydraulics and their applications. Flow measuring instruments in conduits and open channels are also described prior to the listing of basic considerations for eco-friendly design of various water systems.

Chapter 4 discusses various types of qualitative/quantitative hazards arising out of water pollution, floods, landslides, collapse of dams and droughts. It also tabulates the information system requirement for mitigation of various disasters.

Chapter 5, the last among the Chapters, gives the details of traditional/state-of-the-art water conservation/recharge practices in the developing/developed world. Wastewater treatment/reuse by low-cost and localized biological processes are also described. Sustainable development through integrated water management practices (e.g., the construction of facilities which are almost self-sustaining in terms of their freshwater requirement and wastewater generation) are also explained.

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Contents

Chapter 1

Eco-hydrological Background

1.1 ENVIRONMENTAL HYDROLOGY IN GENERAL

Hydrology may be broadly defined as 'water science'. Hydrology is defined as the study of the occurrence and movement of water above, on and below the surface of the earth as well as the properties of water, and water's relationship to the biological and non-biological components of the environment.

The goal of physical hydrology is to explain the phenomena of water-flow in the natural environment by application of physical principles. Solutions to many hydrological problems require an understanding of the dynamics of water motion. The motion of water is through many processes, which occurs in a cyclic manner, through what is known as the *hydrologic cycle* or the *water cycle*.

1.2 HYDROLOGIC CYCLE AND ITS PROCESSES

The fundamental concept of hydrology is the hydrologic cycle: the global-scale, endless re-circulatory process linking water in the atmosphere, on the continents, and in the oceans. This cyclical process is usually thought of in terms of reservoirs (i.e., oceans, atmosphere, etc.) and the volumetric flows of water between them. Within the hydrologic cycle, the dynamic processes of water vapor formation and transport of vapor and liquid in the atmosphere are driven by solar energy, while precipitation and many of the various flows of water at or beneath the earth's surface are driven primarily by gravitational and capillary forces. The area of land in which water flowing across the land surface drains into a particular stream or river and ultimately flows through a single point or outlet on that stream or river is called the *catchment* or the *basin* or the *watershed* in American English. Catchments are delineated on the basis of land-surface topography. The boundary of a catchment is called a *divide* or a *ridge*.

The hydrologic cycle is a continuous process by which water is purified by evaporation and transported from the earth's surface as well as from the ocean surface to the atmosphere and back to the land and oceans. Figure 1.1 is a representation of the hydrologic cycle. All of the physical, chemical and biological processes involving water as it travels through various paths in the atmosphere, over and beneath the earth's surface and through growing plants, are of interest to those who study the hydrologic cycle. There are many pathways that water may take in its continuous cycle of falling as rain or snow and returning to the atmosphere. It may be captured for thousands of years in polar ice caps. It may flow into the rivers and finally to the seas and oceans. It may soak into the soil to be evaporated directly from the soil surface as it dries or be transpired by growing plants. It may percolate through the soil to groundwater reservoirs or aquifers, to be stored or it may flow to wells or springs or back to streams by seepage. The residence time for water may be either a few minutes, or it may take thousands of years. Refer to Table 1.1 for typical values of residence time for different phases.

Table 1.1 Estimate of the World's Waters and their Residence Time

Parameter	*Surface area ($km^2 \times 10^6$)*	*Volume ($km^3 \times 10^6$)*	*Volume (%)*	*Equivalent Depth (m)*[a]	*Residence time*
Oceans and seas	361	1,370	94	2500	~4,000 years
Lakes and reservoirs	1.55	0.13	<0.01	0.25	~10 years
Swamps	<0.1	<0.01	<0.01	0.007	1-10 years
River channels	<0.1	<0.01	<0.01	0.003	~2 weeks
Soil moisture	130	0.07	<0.01	0.13	2 weeks–1 year
Groundwater	130	60	4	120	2 weeks–10,000 years
Ice caps and glaciers	17.8	30	2	60	10-1,000 years
Atmospheric water	504	0.01	<0.01	0.025	~10 days
Biospheric water	<0.1	<0:01	<0.01	0.001	~1 week

[a]Computed assuming uniformly distributed storage over the entire surface of the earth.
[Nace, 1971]

People tap the water cycle for their own uses. Water is diverted temporarily from one part of the cycle by pumping it from the ground or by drawing it from a river or lake. It is used for a variety of activities such as households, businesses and industries, for transporting wastes through sewers, for irrigation of farms and parks and for production of hydroelectric power.

After use, water is returned to another part of cycle. It is either discharged downstream or allowed to seep into the ground. Used water is normally poorer in quality, even after treatment, which often poses a problem for downstream users.

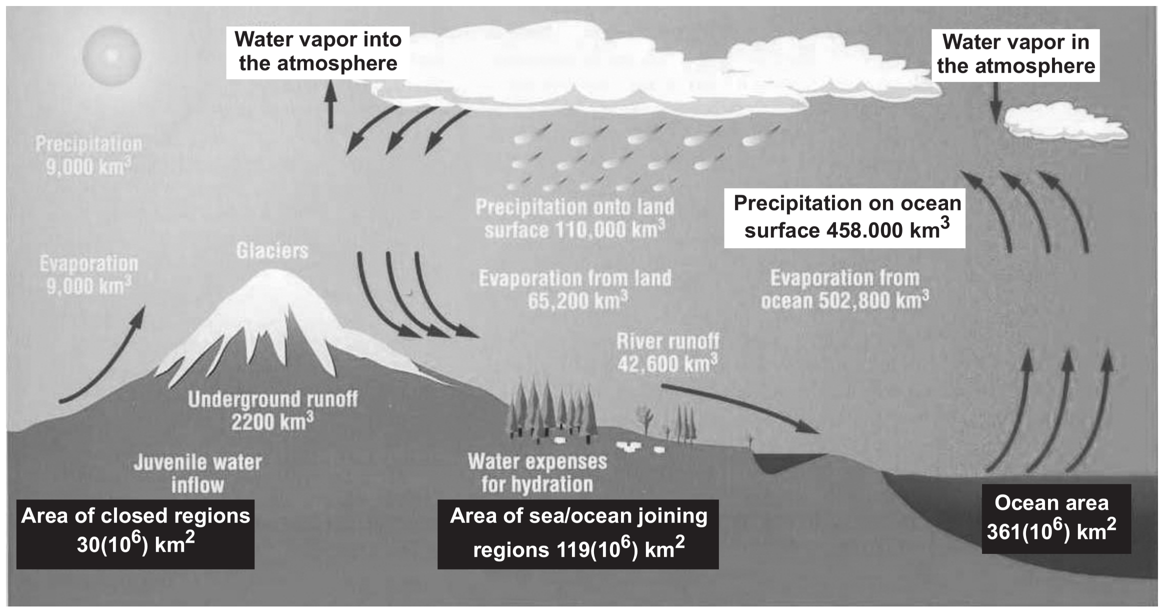

Figure 1.1 Hydrologic Cycle Indicating Global Annual Quantities
[Source: http://www.unesco.org/science/waterday2000/Cycle.htm]

The various processes involved in the hydrologic cycle are as under.

Water vapor enters the atmosphere by *evaporation* and *transpiration*. Evaporation is the process of water in oceans, lakes, rivers, etc. changing into water vapor, while transpiration is the discharging of water vapor into the atmosphere from living vegetation such as leaves, grass, etc. Many a time, evaporation and transpiration are together known as *evapotranspiration*.

Once water vapor enters the atmosphere, it rises and cools. As the water vapor cools, *condensation* or the change from water vapor into liquid water begins to form small drops of water. These small droplets of water are what you look at when you see a cloud. As these droplets bounce around and hit one another, they stick together and make larger drops.

When the drops of water become too heavy to be held up, they fall back to the earth as *precipitation*. Depending on the temperature, it can fall as drizzle, rain, snow, sleet, and many other forms of precipitation.

Once the precipitation hits the ground, it begins to seep into the ground. This process is called *infiltration*. But the soil can hold only a limited quantity of water. And when the ground becomes saturated, the excess water drains into lakes, rivers, oceans, etc. This excess water is called *runoff*. Then the hydrologic cycle starts all over again.

Groundwater is that part of water which is below the ground level. It can be in the form of either storage in the subsurface water bearing strata known as *aquifers* or it can also flow as *groundwater flow* or *subsurface flow* or *subsurface runoff*.

1.2.1 Distribution of Water on the Earth

- Total ~ 1,358 million km^3
- Percentages
  - 97.20% oceans
  - 2.15% ice caps and glaciers
  - 0.65% lakes, streams, groundwater, atmosphere

At its simplest level the water cycle is the balance of evaporation and precipitation. The energy from the sun causes water on the earth to evaporate, this vapor moves through the atmosphere and falls as rain, sleet or snow. The fate of precipitation varies—it may:

1. Never reach the groundwater that is intercepted by buildings or plants, may evaporate rapidly.
2. Land on an impervious surface e.g., pavement or road. Such surface runoff will be channelled away in drains and sewers to rivers and ultimately to the sea or ocean.

3. Reach the soil and permeate it by infiltration. Of this, some will percolate until it reaches an impervious rock layer and form a pool of groundwater. Some will be taken up by the roots of plants and transported to the leaves for use in photosynthesis.

1.2.2 Water Availability and Use

Annual renewable water supply per capita varies greatly around the world: Iceland 670,000 m^3, Canada 112,000 m^3, United States 9000 m^3, Egypt 30 m^3, Kuwait ~0 m^3, Bahrain ~0 m^3.

Water is a crucial environmental resource in many areas of the world. It does not respect political boundaries and consequently disputes exist around the world between neighboring countries about the use and contamination of shared water supplies e.g., in the Middle East, control of the Jordan and the large aquifer under the West Bank and in Turkey, Iraq and Syria over management of the Tigris and Euphrates.

Runoff does not automatically equate to availability because it can be spasmodic e.g., 90% of rain in India falls between June-September. Stable runoff is that which is available year round.

Regional variations in the water cycle are perturbed and sometimes intensified by drought cycles, which put extra pressure on semiarid zones. Undisturbed ecosystems can survive drought but human introduction of non-native animals and plants reduces resistance.

Types of water use are:

- Withdrawal: taken from lake, river or aquifer for non-destructive purpose and returned for re-use.
- Consumption: fraction withdrawn but lost in transmission, evaporation, absorption, and chemical transformation and therefore unavailable for re-use.
- Degradation: change in quality due to contamination, pollution.

Global use of water over the last century has increased by an order of magnitude. Agricultural use is inefficient and highly consumptive with up to 90% of the water never reaching the crops. Industrial use is largely for cooling power plants. If the withdrawn water is not contaminated then it can be re-used. Domestic water use varies according to the availability and the cost.

Many examples exist around the globe of how the intervention of man can disrupt and potentially destroy elements of the water cycle:

- The Ogallala Aquifer lies under 8 states in USA. Once contained 2000 km^3 in porous rock layers (more than all the surface freshwater on the earth). In 1930 average depth was 20 m but it reduced to 3 m by 1987 due to withdrawal for irrigation. Can't be recharged quickly enough. Groundwater depletion can lead to subsidence, sinkholes and salt-water intrusion (coastal areas).

- The Aral Sea was the world's fourth largest lake—shallow and saline. Located in the deserts of Uzbekistan and Kazakhstan (former USSR). It has two river inflows and no outlet but in 1918 it was decided to withdraw water from the inflowing rivers for cotton production. By 1990 the Aral Sea lost 40% of surface area and 66% volume. Fishing villages are now 40 km from the water and salty dust is devastating crops and causing health problems.

Sea level rise over last 100 years of 1-2 mm/yr or more—may reflect the apparent temperature rise due to thermal expansion plus the extraction of groundwater being delivered to the sea and the melting of glaciers. Arctic Sea ice is shrinking and ice shelves on the Antarctic Peninsula are retreating. Most climate models predict a more humid world with evaporation, precipitation and runoff enhanced i.e., a speeding-up of the hydrological cycle. Most temperature change is expected at high latitudes. Observations of increased water vapor in the stratosphere in some areas have been observed with increased rainfall over mid-latitudes and reduced rainfall over N. Africa and the Middle East but historical data is sparse and all must be considered in relation to the long-term cyclical variability in climate.

A hydrologist studies the fundamental transport processes to be able to describe the quantity and quality of water as it moves through the cycle by evaporation, precipitation, stream flow, infiltration, groundwater flow, and other components. The engineering hydrologist, or water resources engineer, is involved in the planning, analysis, design, construction and operation of projects for the control, utilization and management of water resources. Water resources problems are also the concern of meteorologists, oceanographers, geologists, chemists, physicists, biologists, economists, and political scientists, specialists in applied mathematics and computer science, and engineers in several fields.

1.3 PRECIPITATION-RUNOFF-INFILTRATION-EVAPORATION ANALYSIS

1.3.1 Precipitation

Precipitation is most commonly measured daily at a fixed time in the morning, and the measured quantity, expressed as depth in millimeters, is attributed to the previous day. The 'standard' rain gauge varies in orifice diameter and height above the ground in different countries, but most have adopted the World Meteorological Organization (WMO) guidelines of a 150 to 200 cm^2 collector area positioned 30 cm to 1 m above the ground. If rainfall occurs during the morning, and a rain gauge is read during a storm, the total quantity in that storm will be attributed to two separate days. This is an important factor to take into account when assessing the frequency of large falls of rain.

Another type of rain gauge is the recording or autographic gauge which traces amount of rainfall against time on a strip chart. Some newer gauges employ tipping buckets which record, via a magnetic switch, counts or tips in a digital form, on either solid-state or magnetic-tape loggers. These rain gauges are used to analyze the intensity of rainfall and are particularly useful in assessing storm profiles, infiltration rates, surface runoff or rainfall erosivity.

Storage gauges are rain gauges which are left in remote areas for periods of time between visits ranging from 1 week to 1 month. Where evaporation is high, oil can be used to inhibit losses. Calibrated dipsticks are used to measure rainfall between visits. Once again, small solid-state recorders make it possible for storage gauges to be measured at infrequent intervals to give daily rainfalls. The usefulness of all rain gauges is limited by the effectiveness of protection against vandalism.

Radar has been used with some success to measure rainfall in cases where high costs can be justified in order to obtain real time data, as in the case of flood forecasting. Radar has also been used to detect hail formation in the tea-growing area of Kericho in Kenya. Clouds can be seeded to curtail the growth of large hailstones, which can damage a high-value cash crop.

With the ever-increasing quality of satellite imagery it is possible to correlate cloud type, density and thickness with rainfall, using 'ground truth' stations. Once again the cost of such a computer-based exercise has to be weighed against the value of the rainfall data.

Telemetering of rainfall information has been developed in certain remote areas to minimize the transport costs of gathering rainfall data. This, and other remote or automated techniques (including the use of automatic weather stations), is not likely to replace the widespread system of volunteer observers using simple standard rain gauges until realistic monetary values can be placed on sets of reliable rainfall data. However, as technological advances reduce capital installation costs, and as fuel prices continue to rise, the gap between manual and automated systems in terms of total costs may not be so great.

1.3.1.1 Estimation of Rainfall Over an Area

Rainfall over an area is usually estimated from a network of rain gauges. In theory, these gauges should either be placed in a random sampling array, or be set out in a regular or systematic pattern. A combination of sampling techniques, as in a stratified random sampling network, is often used to increase the efficiency of sampling i.e., to decrease the number of gauges required to estimate the mean rainfall with a given precision (McCulloch, 1965).

In practice, gauges are usually placed near roads and permanent settlements for convenience of access, rather than according to a strict sampling array. There is a danger that bias can be introduced into the estimated mean rainfall over an area. This was first pointed out by Thiessen in 1911, who advocated the construction of *Thiessen polygons* (polygons generated around a rain gauge whose sides are the perpendicular bisectors of lines connecting neighboring rain gauges) to give a weighting inversely proportional to the density of rain gauges.

In mountainous areas, for example, fewer gauges tend to be placed in the inaccessible areas which often have the highest rainfall. Careful inspection is needed, therefore, to determine whether the mean is affected by serious bias, and whether this can be corrected using the Thiessen method.

The most rational way of estimating mean rainfall is by constructing *isohyets* (lines joining places of equal rainfall) on a map of the area in question. This takes into account such factors as the increase in rainfall with altitude on the windward side of hills and mountains, the rain shadow on the leeward side of hills and mountains and the aspect of topographic barriers in relation to prevailing winds. The assumption is made that rainfall is a continuous spatial variable i.e., if two rain gauges record 100 mm and 50 mm respectively, there must be some place between the two gauges which have received 75 mm.

Once the isohyets are drawn, mean areal rainfall is calculated by computing the incremental volumes between each pair of isohyets by adding the incremental amounts and dividing it by the total area.

Because of the labor involved in measuring the area between isohyets, several automated techniques have been devised. These include fitting various 'surfaces' to the rainfall data, calculating Thiessen polygon areas automatically or applying finite element techniques (Edwards, 1972; Shaw and Lynn, 1972; Lee et al, 1974; Chidley and Keys, 1970; Hutchinson and Walley, 1972). These techniques are basically designed to accommodate irregularly spaced rain gauges in areas where spatial variation is important, or to provide techniques that can readily be adapted to the computer processing of rainfall data.

In semi-arid areas, the temporal and spatial variability of rainfall is such that rainfall networks are rarely dense enough to reflect adequately either the mean or the variance of areal rainfall. Even where relatively dense networks (e.g., 1 gauge per 700 km^2) are installed as part of research programmes (Edwards et al, 1979), there are difficulties in constructing isohyetal maps, and considerable changes in the seasonal pattern of rainfall can be discerned.

Generally speaking, random sampling networks give a better estimate of the variance of mean rainfall, and systematic sampling networks give a better (i.e., unbiased) estimate of the mean. In some cases, as in network design, the former is more important since it leads to estimates of the precision of the mean or, conversely, to determining the number of gauges necessary for estimating the mean with the required precision. For most general purposes, however, a systematic spatial coverage can be relied upon to give unbiased estimates of the mean.

In this context it is useful to distinguish between an estimate's precision i.e., its repeatability in a sampling sense, and its accuracy, which is a function of both the sampling technique and the estimate of 'true' rainfall at a point.

Rain gauges are of many different heights above the ground. Variations in the cross-sectional area of the gauge from its nominal value, which may be due to poor construction or damage, overexposure of gauges to high wind speeds across the orifice, shelter of the gauge by vegetation, and common observers' mistakes, all contribute to inaccuracies in point rainfall measurements. Normally these are small compared to the seasonal and annual variability of rainfall but, in certain cases, these potential sources of error have to be taken into account. Such cases include the measurement of rainfall above forest canopies and in areas of high wind speed.

1.3.1.2 Frequency Distribution of Annual, Monthly and Daily Rainfall

Agriculturalists and pastoralists require statements concerning the reliability of rainfall. A convenient means of making such statements is provided by *confidence limits*. These are defined as the estimates of the risk of obtaining values for a given statistic that lie outside prescribed limits (Manning, 1956). The limits commonly chosen are 9:1 and 4:1. With 9:1 limits a figure outside the limits is to be expected once in 10 occasions, and half of these occasions (i.e., one in 20) are to be expected below the lower limit and half above the upper limit. Thus the 9:1 lower confidence limit of annual rainfall would represent the level of rainfall which is expected not to be reached once in 20 years. Similarly, the 4:1 lower confidence limit is expected not to be reached once in 10 years.

In order to establish values for such limits, it is necessary to assume a theoretical frequency distribution to which the sample record of data can be said to apply. Manning (1956) assumed that the distribution of annual rainfall in Uganda was statistically normal. Jackson (1977) has stressed that annual rainfall distributions are markedly 'skewed' in semi-arid areas and the assumption of a normal frequency distribution for such areas is inappropriate. Brooks and Carruthers (1953) make general statements that three-yearly rainfall totals are normally distributed, that annual rainfall is slightly skewed, that monthly rainfall is positively skewed and leptokurtic, and that daily rainfall is 'J'-shaped, bounded at zero. They go on to suggest that empirical distributions be used, such as log-normal for monthly rainfall, and an exponential curve (or a similar one) fitted to cumulative frequencies for daily rainfall.

For annual rainfall series which exhibit slight skewness and kurtosis, Brooks and Carruthers (1953) suggest that adjusting the normal distribution is easier and more appropriate than using the Pearson system of frequency curves. As the degree of skewness and kurtosis increases, log-normal transformations should be used.

These comments apply equally well to tropical rainfall where annual totals exceed certain amounts. For example, Gregory (1969) suggests that normality is a reasonable assumption where the annual rainfall is more than 750 mm. Kenworthy and Glover (1958) suggest that in Kenya normality can be assumed only for wet-season rainfall. Gommes and Houssiau (1982) state that rainfall distribution is markedly skewed in most Tanzanian stations.

Inspection of the actual frequency distribution, at given stations, and simple tests for normality can quickly establish whether or not the normal distribution can be used. Such tests include the comparison of the number of events deviating from the mean by one, two or three standard deviations with the theoretical probability integral. If not, a suitable transformation must be chosen before confidence limits or the probabilities of receiving certain amounts of rainfall can be calculated.

Maps have been prepared for some regions, particularly East Africa (East African Meteorological Department, 1961; Gregory, 1969), showing the annual rainfall likely to be equalled or exceeded in 80% of years. These are extremely useful for planning purposes although, as Jackson (1977) points out, these refer to average occurrences over a long period of years. Statements such as 'the rainfall likely to be equalled or exceeded in 4 out of 5 years' must be qualified by 'on average' to indicate that it does not rule out the occurrence of, say, 3 years of 'drought' rainfall in a row.

1.3.1.3 Frequency Distribution of Extreme Values

When dealing with the frequency distribution of maxima or minima, it is necessary to use other empirical frequency distributions which give a more satisfactory fit to the observed data. There are no rigid rules governing which type of distribution is most appropriate to a particular case, and a variety of empirical frequency or probability distributions are available in standard statistical textbooks (Table 1.2). As a general guide, extreme distributions are concerned with the exact form of the 'tail' of the frequency distribution. Because such occurrences are rare events, it is unusual to have a sufficiently long record for the shape of the asymptotic part of the curve to be defined with any certainty. Brooks and Carruthers (1953), for example, feel that the Gumbel distribution, which is commonly used in flood frequency prediction, tends to underestimate the magnitude of the rare rainfall events.

Table 1.2 Common Types of Frequency Distributions used for Hydrological Events

	Type	*Characteristics*	*Example*
1.	Binomial	Discrete events in two categories	Number of 'dry' months in each year
2.	Poisson	'J'-shaped distribution of discrete events; possible number of occurrences very small	Frequency of heavy rainstorms
3.	Normal	Symmetrical, bell-shaped continuous distribution (Gaussian)	Annual rainfall in wet regions
4.	Adjusted normal	Like 'normal distribution' but slightly skew	Annual rainfall
5.	Log-normal	Positively skew and leptokurtic	Annual rainfall in semi-arid areas, monthly rainfall
6.	Pearson Type I	Bounded both ends, bell or 'J'-shaped, skew	Monthly rainfall with very dry months
7.	Pearson Type III	Bounded one end, bell or 'J'-shaped, skew (log-normal is a special case)	Frequencies of wind speed
8.	Extreme distribution Type I	Asymptotic, unbounded (Gumbel)	Flood frequency, rainfall intensity/frequency
9.	Extreme distribution Type III	Asymptotic, bounded at a minimum value	Drought frequency

In most applications it is usual to linearize the distribution by calculating cumulative frequencies and then plotting on double logarithmic versus linear graph paper (extreme probability graph paper). If the actual distribution is close to the theoretical distribution postulated by Gumbel, the plotted points approximate to a straight line. An example of this is shown in

Figure 1.2 which demonstrates how a cumulative distribution can be linearized by the use of special probability paper (Ven te Chow, 1964). The same author points out that the abscissae and ordinates are normally reversed to show the probabilities as abscissae. Figure 1.3 is an example of the more usual presentation of data plotted on extreme probability paper. In this case, it is seen that the maximum daily rainfalls for June, in Lagos, Nigeria closely fit the Gumbel distribution.

In this context it is useful to mention that most analyses of frequencies of rainfall or other hydrological events are expressed in terms of the *recurrence interval* of an event of given magnitude. The average interval of time, within which the magnitude of the event will be equalled or exceeded, is known as the *recurrence interval* or *return period* which is the reciprocal of *frequency*. It is common, therefore, to refer to the 10-year or the 100-year flood.

The recurrence interval (N) should not be thought of as the actual time interval between events of similar magnitude. It means that in a long period, say 10,000 years, there will be 10,000/N events equal or greater than the N-year event. It is possible, but not very probable, that these events will occur in consecutive years (Linsley and Franzini, 1964).

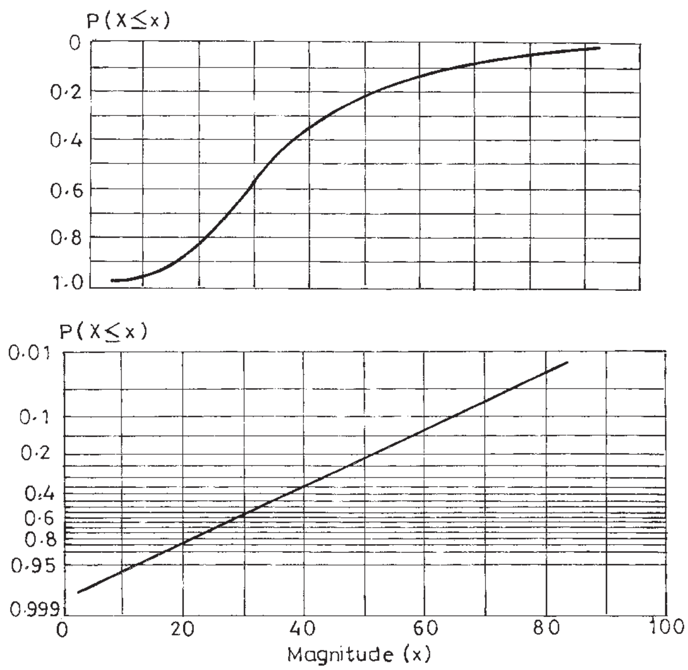

Figure 1.2 Linearization of a Statistical Distribution. Cumulative Probability Curve Plotted on Rectangular Coordinates.

[Source: Ven te Chow (1964)]

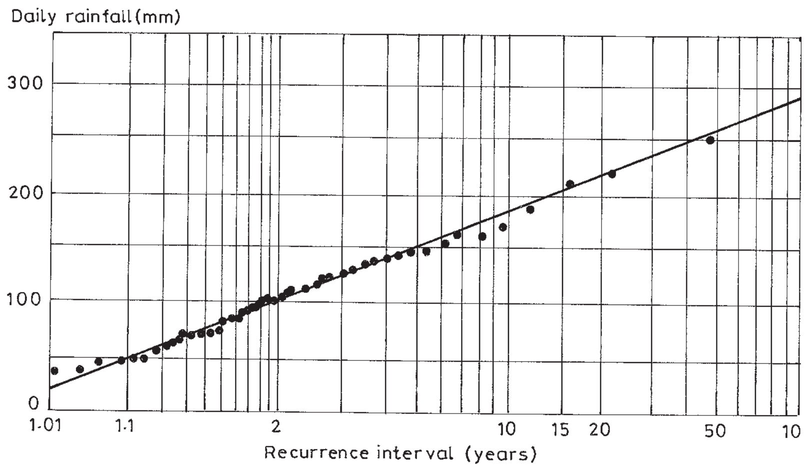

Figure 1.3 Gumbel Distribution for Maximum Daily Rainfall in Lagos, Nigeria for June During 1915-1926.
[Source: Ayoade (1977)]

Most of the record falls of rain are associated with intense tropical cyclones. Figure 1.4 shows the world's highest recorded rainfalls. The totals recorded at Cilaos (Reunion) were the result of an intense tropical cyclone in 1952, which was funnelled up a steep valley rising to 3000 m (Lockwood 1972). The previous world record for 24 hours was also due to the combination of tropical cyclone and land relief. On that occasion, the maximum rainfall occurred at Baguis in the Philippines following the passage of a typhoon in 1911. Less intense, but still remarkable falls of rain commented upon by Jackson (1977) are also shown on the graph.

Rainfall maxima such as those in Figure 1.4 only give an idea of the expected magnitude of the highest falls, with very long recurrence intervals. For general design purposes, statements about amount, duration and frequency of rainfall are required in order to compare the risk of failure of spillways, culverts and bridges against economic criteria. Such analyses can also be applied to soil erosion problems.

For a given duration, rainfall events can be ranked in either a partial duration series or an extreme value series. Thus, for daily rainfall, events can be selected so that their magnitudes are greater than a certain base value, or the maximum rainfall in any year can be chosen. Either series can be plotted on probability paper to yield recurrence intervals or return periods for a given magnitude. The difference between the two series is that the partial duration series may include several events which occur close together in 1 year. For practical purposes, the two series do not differ much, except in the

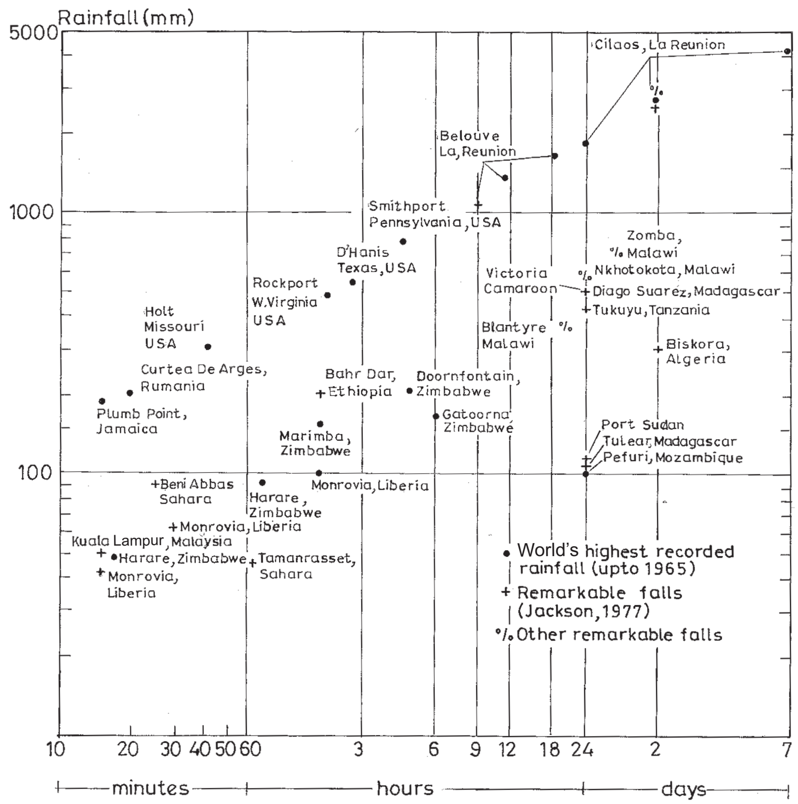

Figure 1.4 World's Highest Recorded Rainfall and Other Remarkable Falls.

[Source: http://www.fao.org/Wairdocs/ILRI/x5524E/x5524e0h.gif]

values of low magnitude (Ven te Chow, 1964). Alternatively, where information from recording rain gauges is available, rainfall intensity can be plotted against duration, and intensity-duration frequency curves can be constructed for different return periods.

1.3.1.4 Occult Precipitation

On mountain masses within the tropics, the summits are frequently covered in cloud in locations where the orographic uplift of moist air is sufficient to cause condensation. If the upper slopes are covered in forest, the surfaces of the leaves and branches of trees act as collectors for condensed water vapor. This phenomenon is referred to as *occult precipitation*, because it does not relate to any rain-making process in the cloud itself and rain gauges placed outside the forest area will not record any rainfall.

In southern and southwestern Africa, Marloth in 1904 and Nagel (1962) have shown that mist condensation or occult precipitation may contribute between 40 and 94% of the total precipitation on high ground. More recently, Edwards et al (1979) made some preliminary measurements on Mt. Kulal in northern Kenya and concluded that occult precipitation could be a significant addition to the groundwater store.

This mechanism may be important as a contribution to groundwater recharge and as a means of sustaining perennial basal springs on mountains in semi-arid areas. On the other hand, high rates of evaporation of intercepted water, and the deeper rooting zone of trees compared to shorter vegetation, may mean that any additional infiltrated water from occult precipitation will be rapidly used up by the trees.

Persistent and subjective reports of springs drying up following deforestation, and flowing again following afforestation, point to the value of investigating the detailed water balance of mountains in semi-arid areas. Unfortunately, the water balances are likely to be complex, with both infiltration capacities and rates of input of precipitation varying between forested and non-forested areas. Evaporation and transpiration rates are also likely to differ significantly and, because these components would need to be measured directly, projects aiming to achieve detailed water balances would require careful instrumentation.

1.3.2 Runoff

The volume of storm runoff depends to a large extent on the antecedent soil moisture conditions and the intensity of rainfall. Typical annual runoff volumes in semi-arid areas, expressed as a percentage of rainfall or *runoff coefficient*, would be less than 10%, but individual storms can produce much higher runoff volumes (~70%) when rain falls in an already saturated catchment. Runoff varies widely according to the seasonal distribution of rainfall, catchment characteristics (shape, size, steepness), vegetation type and density, in addition to the two basic criteria mentioned above. Generally speaking, the wide variation of annual totals makes it difficult to predict total catchment yield from rainfall alone. Most attempts to regress runoff against rainfall lead to unacceptable scatter.

While total runoff or stream flow gives the yield of water from a particular catchment in a particular year, it is closely dependent upon variations in annual rainfall. As in the case of rainfall, a set of runoff data can be subjected to a frequency analysis and statements can be made about the probability of occurrence of certain values.

More usually, however, interest is centered on the extremes of exceptionally high or low annual runoff events and their duration. In these cases, partial duration series or annual maximum (or minimum) series plotted on extreme probability paper will give the recurrence intervals of particular events.

To obtain the duration of annual flows above or below certain limits, an entirely different technique is required. This relies on rainfall being a stochastic process, where the annual rainfall in a particular year is independent of the rainfall in preceding years. Although irregular trends and spells of wetter-than-average or drier-than-average rainfall are commonplace, most attempts to discern reproducible cycles or harmonic patterns in rainfall have failed. In the absence of demonstrable cycles there is no reason to assume that the irregular patterns are other than those, which could be expected to occur in an entirely random series from time to time.

It is possible to extend a data set having characteristic statistics of mean and variance by generating a similar sequence of random values. This can be repeated many times and the actual occurrence of runs of values above or below certain limits can be analyzed. A good example of this technique in practical use is described by Kidd (1983), who used it to generate synthetic or theoretical inflows into Lake Malawi in order to predict outflow down the Shire River and the effects of controlling outflow with a barrage. Table 1.3 lists the runoff coefficient for the rivers around lake Malawi.

Often, as in the case of designing small water supplies, the seasonal variations in daily flow are important. The best way to represent these is by means of a *flow-duration curve*. This is a cumulative frequency curve of daily flows, expressed as a percentage over as long a period as possible. Thus, it is possible to estimate the percentage of time during which a given flow is equalled or exceeded.

In the case of low flows duration of flow is particularly important. It is often necessary to know what the minimum flow would be over a certain time with a given level of occurrence. Thus for a water supply scheme, 30 days might be the absolute maximum that either live stock or people could survive on storage. The 5-year, 30-day minimum flow can be compared with the demand to see what supplementary measures (boreholes, shallow wells) are necessary, on average, in a 5-year period.

This statistic and other comparable statistics (e.g., 10-year, 60-day minimum flow) can be calculated by a frequency analysis of overlapping 'n'-day flows in a given runoff data. This is best performed on a computer because of the amount of data involved. Where a long dry season occurs, as

Table 1.3 Annual Runoff Coefficient – Lake Malawi, Africa

River	*Mean annual rainfall (mm)*	*Mean annual runoff (mm)*	*Runoff coefficient (%)*
Ruo	1280	395	31
Kwakwazi	1240	332	27
Likangala	1430	499	35
Domasi	1730	882	51
Naisi	1280	312	24
Linthipe	880	133	15
Lilongwe	930	155	17
Lingadzi	810	82	10
Bua	900	83	9
Bua	900	73	8
Dwangwa	740	37	5
Luweya	1480	500	34
Luchelemu	1090	260	24
Lunyangwa	1210	232	19
N. Rumphi	1320	661	50
Kambwiya	1370	625	46
N. Rukuru	910	227	25

[Source: http://www.fao.org/Wairdocs/ILRI/x5524E/x5524e04.htm#3.5%20runoff]

in much of tropical Africa, a visual inspection of the recession curves of the streams is usually sufficient to gain some idea of the frequency of very low flows of specified duration.

It is necessary to know the peak flow or flood for the design of any structure intended to pass a given volume of water per unit time (e.g., spillways on dams, culverts, canals and bridges). Many empirical and analytical techniques are available to predict floods. It is outside the scope of this book to discuss more than the general principles. Readers interested in this topic will find a comprehensive treatment of the subject in NERC (1975), in spite of the direct relevance of this report to British conditions. The proceedings of the symposium on 'Flood hydrology' (TRRL, 1977) deal specifically with African conditions, and apply similar techniques of flood prediction to both urban and rural catchments.

Basically, a choice can be made between empirical techniques based on catchment characteristics, statistical techniques (where there is an abundance of reliable data), and unit-hydrograph techniques (which require some knowledge of rainfall characteristics, soil types, channel slope and shape of the unit hydrograph).

In many areas, there are insufficient streamflow records for statistical techniques to be applied with confidence. Usually statements are required about ungauged catchments where a large degree of uncertainty is bound to arise. Choice of method depends to a large extent, therefore, on the amount and type of data available. It is still common to find the so-called '*rational method*' being applied where sufficient data for other methods are lacking (Prabhakar, 1977). In this method, maximum runoff (Q_{max}) is related to average rainfall intensity (I) and catchment area (A) by means of an empirical runoff coefficient (K):

$$Q_{max} = K \cdot I \cdot A \tag{1.1}$$

Values of K are derived locally, often by trial and error, and values are given in Table 1.4.

Table 1.4 Empirical Values of the Factor K in the Rational Formula

Type of catchment	*Empirical values of factor K*
Rocky and impermeable	0.30 to 1.00
Slightly permeable, bare	0.60 to 0.80
Slightly permeable, partly cultivated or covered with vegetation	0.40 to 0.60
Cultivated absorbent soil	0.30 to 0.40
Sandy absorbent soil	0.20 to 0.30
Heavy forest	0.10 to 0.20

Source: Prabhakar (l977).

Another method, used when data are lacking, is the '*envelope curve*' method in which recorded peak flows are plotted against catchment area. Although a remarkable degree of conformity is often found, it is not possible with this method to relate magnitude to frequency of occurrence; but from experience, it is expected to give approximately the 25-year flood.

Implicit in many analyses of streamflow is a division between *surface runoff*, or *storm runoff*, and *base flow*. As the name implies, surface runoff is the direct runoff from the soil surface to the stream course, concentrated by the shape of the catchment and the drainage network into a flood wave, which reaches a maximum and then attenuates as it travels down the main river. Base flow, on the other hand, is conventionally taken as the contribution to streamflow from groundwater. After the passage of the flood wave, the catchment slowly drains until the flow in the river is related to the amount of water held in storage within the catchment. The '*recession curve*', the slowly falling limb of the hydrograph, therefore, reflects the rate of release of water from the groundwater store, which is itself dependent upon a combination of hydraulic head and saturated permeability of the catchment. The exponential shape of a recession curve allows 'recession constants' to be calculated. These in turn can be used to predict the amount of water in the groundwater store, since the physical process of releasing water from storage can be simulated by a *linear reservoir* (in systems terminology) whose outflow is directly related to storage:

$$Q_t = Q_o K_r^{\,t} \tag{1.2}$$

where Q_t = flow at time t after Q_o, and K_r = recession constant.

The storage (S_t) remaining in a basin at time t is given by:

$$S_t = -Q_t/\ln K_r \tag{1.3}$$

The recession constant can be used, therefore, both to separate stormflow and baseflow (by extrapolating backwards in time from the recession limb of the hydrograph) and to estimate the groundwater storage at a given time. The latter technique is often used in water balance calculations, as will be seen later.

1.3.2.1 Measurement of Streamflow

Although hydrometric networks have improved over the past 30 years, the majority of gauging stations are still on large rivers or in areas of high agricultural potential, where information is required for water supply and/or irrigation.

In the rangelands not only are networks sparse, but the preponderance of shifting controls (e.g., sandy river beds), coupled with infrequent visits, lowers the reliability of the streamflow records.

In ungauged catchments, which are often of very small size, it is frequently necessary to estimate either storm runoff or streamflow. Once again there is a choice of methods, depending on the type of information required and the availability of data from nearby stations within the same hydrological region. These range from the statistical, analytical and empirical techniques referred to in the last section to mathematical models of various degrees of complexity, which are dealt with later in this chapter. However, the actual measurement of streamflow is better than the most sophisticated indirect methods, although usually a combination of the two is used to derive estimates of frequency or reliability.

Measuring the flow in a river is straightforward if the river is shallow enough for wading. The basic problem is that of obtaining a unique relationship between height of the water above a datum and discharge. Weirs and flumes are structures, which stabilize the channel section to give this unique relationship. For very small streams with a *discharge* (i. e., flow per unit time) up to 0.12 cumecs, portable "v"-notch weirs can often be installed (British Standards Institution, 1965), which will measure discharge for a given depth of flow over the weir. Where the benefits to be accrued do not warrant the construction of weirs or flumes, the natural channel has to be 'calibrated'. This is achieved by making a series of measurements of discharge, using a current meter, at different '*stages*' (i.e. at different heights of water above the zero-flow datum). These values can be plotted on a graph of discharge against stage, to give a smooth curve (the 'stage-discharge' or 'rating' curve). This curve can be extended using the Manning formula, which relates discharge to the slope of the channel and its conveyance (Ven te Chow, 1964). Once the rating curve is established, continuous or periodic measurement of a stage can be readily converted into discharge.

The choice of channel section is important in order to ensure, as far as possible, that the flow remains uniform, and that a '*control*' governs the stage-discharge relationship throughout its range by eliminating the effects of downstream channel features. Such controls can be rock bars or constrictions of the stream channel, which are characterized by having pools or smooth reaches upstream. Where such natural constrictions do not exist, long, straight reaches with stable beds should be sought. It is possible with the latter, however, that the control may be drowned out at high flows causing a change in the stage-discharge relationship.

Where a river is carrying a lot of sediment, changes in the channel cross-section will occur from time to time. These will affect the stage-discharge relationship, and will call for frequent measurements of discharge with a current meter to ascertain what error is being introduced into the discharge calculations by assuming a single rating curve. Often several rating curves will have to be used where a channel is eroding or silting up rapidly.

For a full discussion of the techniques for measuring streamflow the Norwegian Agency for International Development (NORAD) has produced a series of manuals on operational hydrology (Tilren, 1979) for the use of the Tanzanian Ministry of Water, Energy and Minerals. Other users would find these manuals full of sound practical information, which could be applied throughout tropical regions of the world.

1.3.2.2 Application of Hydrology in the National Weather Service

An example for the application of hydrology in weather prediction in USA can be cited here. There are 13 River Forecast Centers in the USA that are responsible for forecasting the water levels in the nation's major rivers and their tributaries. Each River Forecast Center is assigned a certain area of responsibility. Figure 1.5 shows the area of coverage of LMRFC—Lower Mississippi River Forecast Center. This Center covers an area of about 542,000 km^2.

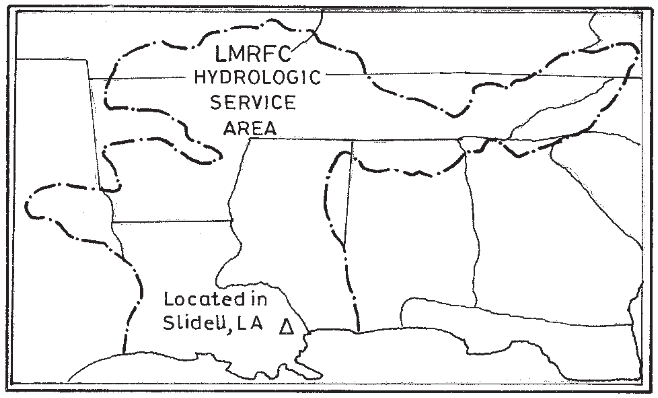

Figure 1.5 The Coverage Area of Lower Mississippi River Forecast Center (LMRFC) in USA.

[Source: http://www.srh.noaa.gov/lmrfc/about/#intro]

The following is a simplistic view of how to make a river stage forecast by using the basic steps below:

1. Data collection of stages, reservoir flows and mean area precipitation and future rainfall.
2. Data entry into the database.
3. Data pre-processing—Quality control of the data to make sure figures are correct.
4. Input into a hydrologic model. This contains all pertinent information at each forecast point, such as the present moisture in the top and bottom layers of the soil, to the climatology of how it reacts to rainfall.

The hydrologist is then able to obtain a forecast of the discharge and then relate this to a stage. River stage forecasts are made daily and are sent out to the Weather Service Forecast Offices. The River Forecast Centers (RFCs) make 3-day and also 5-day forecasts. Long-range forecast for 7 days and up to 30 days are made along the Mississippi River weekly by the LMRFC. When flooding is imminent, forecast is made for the time and height of the crest of the stage and also when the river is expected to go above flood stage. Another product sent out by the RFCs for use by the Weather Offices is the

Flash Flood Guidance. The Flash Flood Guidance is an indicator of the amount of rainfall required to produce flooding.

A *storm hydrograph* is a graph of river stage or discharge at a gauging site versus time. But before going into more detail of the hydrograph, let's define some terms. A storm hydrograph is made up of two components, viz., the base flow and surface runoff. Base flow is that rainfall which seeps (infiltrates) into the soil and moves laterally to the stream channel, reaching the stream channel after several days or more. Surface runoff consists of the rainfall and/or snowmelt, which travels overland to the stream channel plus rainfall directly deposited on the channel. Surface runoff reaches the stream channel quickly and is the main contributor of the total flow during flooding. Some main factors affecting runoff are:

1. Rainfall amount—High rainfall amounts produce more runoff than low rainfall amounts.
2. Rainfall intensity—For the same total amounts of rain, more runoff will occur with rain falling in short periods of time as opposed to rain falling in a longer period of time. For example, one inch of rain falling in 30 minutes will produce more runoff than one inch of rain falling in 24 hours. The lower runoff amounts are due to the water having a longer time to seep (infiltrate) into the ground.
3. Soil type—More runoff will occur with clay soils while sandy soils are able to absorb more rainfall.
4. Soil moisture—When the top layer of soil is moist, there will be more runoff than if the soil moisture content is low.
5. Vegetation—Vegetative cover may slow the runoff from rainfall. As vegetation takes in water, the runoff is retarded.
6. Topography—Runoff varies as the terrain varies. A mountainous terrain will have a faster runoff rate than one of a flat terrain.
7. State of ground—Rainfall over frozen ground produces more runoff than rainfall over non-frozen ground. Frozen ground is more impervious to rainfall.

1.3.2.3 Flooding

Flooding kills more people and causes more damage per year on average than any other natural disaster including hurricanes and tornadoes. There are many different forms of flooding, but only the most common and most destructive kinds will be covered here briefly. They are:

1. *Rainfall Flooding*: This is the most common form of flooding. When it rains enough to cause runoff, the runoff flows into the rivers and the

water levels begin to rise. The river will rise above its banks and flood the surrounding area when more water than the river can hold flows into it. This type of flooding can happen within less than an hour after the rain begins (usually in mountainous regions) or even days after the rain has stopped (in flat regions). It can also happen in a place where there was no rain at all! The rain can fall over an area far upstream and be carried downstream, therefore flooding areas where no rain had fallen.

2. *Snow Melt Flooding*: This type of flooding usually occurs in the cold regions, but water from melted snow can influence river heights all the way down to the delta. When temperatures begin to rise in the spring and there is an excessive amount of snow on the ground, the snow begins to melt and flow into the rivers. This type of flooding is especially dangerous when temperatures rise well above freezing (i.e., 10°C and above) combined with additional rainfall.
3. *Ice Jam Flooding*: This type of flooding only happens in cold climates. Thick ice forms on the surface of rivers during periods of continuous cold weather. When the water level rises quickly underneath the ice, the ice heaves up, cracks, and is carried down the river. The large chunks of ice get stuck under bridges or in sharp bends in the river. The ice blocks the flow of water and the water flows over its banks and floods the surrounding area. This type of flooding is usually confined to smaller areas.
4. *Dam Break Flooding*: This is usually the most destructive type of flooding. Dams can hold back literally millions of cubic meters of water which form huge man-made lakes that can be used for many things such as recreation (boating, swimming, fishing, etc.), flood management (holding excess water which would normally go down river and flood), and city water management (holding water for people to use for drinking, showering, etc.). But when the dams are not constructed or maintained correctly, it creates the potential for disaster. When a weakened dam breaks, it sends a huge wall of water down the river which can destroy entire towns in a matter of minutes. Fortunately, this very destructive type of flooding is also the least common.

1.3.3 Interception and Infiltration

Foresters have always observed the phenomena of interception and throughfall, but few have been prepared to acknowledge that these components were of great significance in the context of the hydrologic cycle. A recent study has established that interception can play a major role in the water balance of catchments where the aerodynamic component of the energy balance is large relative to net radiation. In the tropics this is not likely to be the case, but there are indications that, even in forests lying not too far from the equator, evaporation of intercepted water can be a significant feature.

Intercepted moisture, stored in the canopy, is the first component of the hydrological cycle to be lost directly back to the atmosphere. In areas of high wind speed, with aerodynamically 'rough' canopies, this loss can be very rapid and, in areas where the canopy is frequently wetted, the total quantity of intercepted water lost by evaporation can be a significant proportion of the total rainfall. Interception of raindrops by canopies is also a major factor in reducing soil erosion. This has an indirect effect on the hydrological cycle, in that, by conserving surface soil, infiltration is maintained.

In areas of shorter vegetation interception storage is likely to be small, and the rate of loss may not exceed the potential evaporation rate. Thus in rangelands, interception storage is unlikely to be a measurable quantity in the water balance. Many dry-season grazing areas, however, depend on perennial springs for water supply. These springs emanate from hillsides covered in vegetation, where interception protects the slopes through which the springs are being recharged. Throughfall, stem flow and drip (including the occult precipitation described earlier) form the precipitation input. In dense canopies throughfall is of minor importance, and hence the rate at which water is received by the soil surface is within its infiltration capacity. Surface runoff is practically nil on the heavily forested slopes, and deep percolation is often rapid, through fractured or weathered bedrock. These advantages in terms of recharge, however, have to be offset against the transpiration of deep-rooted, perennial vegetation. This tends to produce moisture deficits in the root zone. These deficits inhibit deep infiltration until sufficient rain falls to saturate the root zone.

Over most of the rangelands in tropical Africa infiltration is the key process in the hydrological cycle. *Infiltration capacity* as termed by Horton or *infiltrability* (Hillel, 1971) which is defined as the rate at which water can enter the soil. Horton recognized both a maximum and a minimum infiltration capacity of soils; the maximum being at the onset of rainfall with the capacity decreasing as the impact of raindrops changes the surface structure of the soil.

When the rate of rainfall exceeds the infiltration capacity of the soil, the result is surface runoff, unless the surface water is stored in depressions (surface or depression storage). Thus, management for soil conservation is directed both at maintaining high infiltration rates by preserving a good surface cover, and at increasing surface storage by pasture furrows, cut-off ditches, and range-pitting or water-spreading.

One of the additional problems in grazing areas is that of '*puddling*'. Puddling is the structural change associated with mechanical stress while soils are in a moist condition, and results in the destruction of large pores in the soil through which water percolates. This has the effect of decreasing infiltration capacity and increasing surface runoff. Further mechanical sorting results, together with the removal of organic material and, often, surface sealing. In fact, it was found that from different types of land use in a catchment in Machakos, Kenya the greatest loss of soil was from degraded grazing areas. Much of this can be attributed to loss of vegetation from overgrazing and the downward spiral of poor vegetation leading to high rates of soil loss and surface runoff.

With some soils, subsoil permeability may be the limiting factor in determining infiltration rates. While this points to surface management being less important, experimental evidence shows that the compaction of well structured surface layers in tropical forest soils, such as that following clearing, causes infiltration capacity to drop by half.

Although infiltration can be seen to be a key process in the cycle, very few experimental data are available to quantify infiltration capacity. It is known to be spatially highly variable, and for this reason isolated measurements are of little practical benefit. Indirect evidence of the improvement of infiltration, following the re-establishment of grass cover, can be found in Edwards and Blackie (1981), who report the results of the Atumatak catchment experiment in Karamoja, Uganda from 1959 to 1970. Two adjacent degraded catchments were chosen for this experiment. One was fenced, cleared of secondary bush and subjected to a controlled grazing-density scheme. The other continued under uncontrolled grazing conditions. Soil moisture tension blocks were installed at a number of sites and, from an analysis of the measurements, it was clear that, in the post-clearing phase, infiltration penetrated down to a depth of 60 cm at most sites. A corollary to the recovery in infiltration in the cleared catchment was the reduction in storm runoff to half that in the 'control' catchment - a striking example of rapid recovery following a modest management programme.

1.3.3.1 Groundwater

The infiltration characteristics of the land surface and the rainfall intensity and duration determine the rate at which the soil moisture store is replenished. After consumptive use by vegetation, a small proportion drains under gravity to the groundwater store. The occurrence of subsurface water can vary according to soil type, the nature of the underlying parent material and the depth of weathering. A classification of subsurface water is given in Figure 1.6. Water is held in each zone as a result of gravitational surface tension and chemical forces. There are no sharp boundaries, except at the capillary fringe in coarse-grained sediments. The water table is, in fact, a theoretical surface, and can be demonstrated approximately by the level of water in wells, which penetrate the saturated zone. The water table can be defined as the level at which the fluid pressure of the pores, in a porous medium, is exactly atmospheric. It can exist in the rock pores to very great depths of about 3000 m, but in dense rock the pores are not interconnected and the water will not migrate.

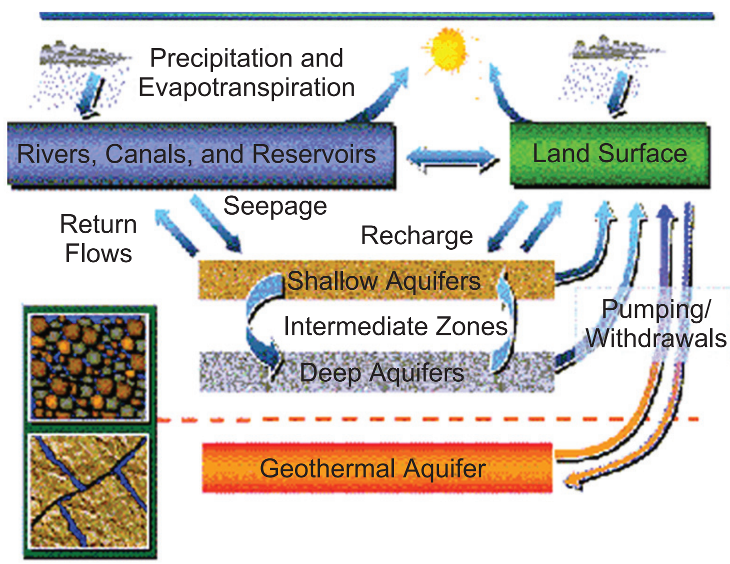

Figure 1.6 Diagram of Aquifers in Relation to Hydrologic Cycle.
[Source: Anon., 2003a]

Only a small proportion of the zone of saturation will yield water to wells. The water-bearing portions are called aquifers. Many types of formation can serve as aquifers, a key requirement being the ability to store water in pores. Table 1.5 gives a typical range of values of porosity, and it is clear that the unconsolidated deposits (chiefly sands and gravels) are the most important aquifers. Pores in the silts and clays are too small. Volcanic rocks are often good aquifers because of the many types of openings, which contribute to their permeability. Igneous and metamorphic rocks are generally regarded as poor aquifers but the Precambrian basement rocks are near the surface and are deeply weathered. As in the case these form shallow aquifers of low yield (0.5 to 1 l/sec), sufficient for small domestic supplies and stock watering. The areal extent of the basement complex rocks on the old erosion surfaces makes this type of aquifer extremely important in the context of tropical Africa. Joints and fractures in the crystalline basement hold some water, but storage is usually quite small and recharge is from the surface weathered zone. Therefore, yields tend to be less than or equal to those from the shallow surface aquifers.

Table 1.5 Typical Porosity Values for Different Formations

Aquifer type	*Porosity (%)*
Unconsolidated deposits	
Gravel	25-40
Sand	25-50
Silt	35-50
Clay	40-70
Rocks	
Fractured basalt	5-50
Karst limestone	5-50
Sandstone	5-30
Limestone, dolomite	0-20
Shale	0-10
Fractured crystalline rock	0-10
Dense crystalline rock	0-5

[Source: http://www.fao.org/Wairdocs/ILRI/x5524e04.htm#3.4%20groundwater]

Aquifers may be classified as confined and unconfined, according to whether or not the water is separated from the atmosphere by impermeable material. Confined aquifers often give rise to artesian wells, where the pressure within the aquifer is sufficient to produce flowing wells at the surface. Unconfined aquifers, of course, have water whose upper surface is at atmospheric pressure. The term semi-confined is used for the intermediate condition, where the confining layer is not completely impermeable. Often lenses of unconfined water are encountered above the water table, held by isolated layers of impermeable material. These are termed perched aquifers and the upper surface is a perched water table. Figure 1.6 gives a diagram of aquifers in relation to hydrologic cycle.

Groundwater may be discharged at the surface or into bodies of surface water. Springs are the most noticeable manifestation of this, but seepage into rivers and lakes is also an important part of the hydrological cycle. Well digging or borehole drilling induces artificial discharge. The quantity of discharge is a function of the porous medium, which is related to the size and interconnection of the pores. The storage term for unconfined aquifers is the 'specific yield' defined as the volume of water released from storage per unit surface area of aquifer, per unit decline in water table. The storage coefficient in confined aquifers is the '*storativity*'—defined as the volume of water released per unit surface area of aquifer, per unit decline in the component of hydraulic head normal to that surface (Freeze and Cherry, 1979).

In the context of the hydrological cycle, groundwater flow passes from recharge areas to discharge areas. Fluctuations in the water table introduce transient effects in the flow system. However, it can often be simulated as a steady-state system if the fluctuations in water table are small in comparison with the total vertical thickness of the system.

Discharge areas in semi-arid climates can be mapped by the direct field observation of springs and lines of seepage, or by the occurrence of phreatophytes and other distinctive vegetative patterns. An analysis of the lithology, topography and existing borehole data will also give information on the recharge-discharge regime and allow the application of steady-state water balance equations to the surface and subsurface components within the recharge and discharge areas.

This approach is often applied to determine the recharge potential within a catchment area and, although livestock requirements are generally very small in comparison with the recharge components, it is a useful exercise to balance all the measured components in order to determine gross errors, or to assess the possible effects of a management programme.

1.3.4 Evaporation and Transpiration

The problem of measuring evaporation from open water surfaces, and transpiration from different types of vegetation, has been a central problem in hydrology for many years. In terms of the hydrological cycle and the water balance, evaporation and transpiration make up the second largest component. Errors in estimating evaporative loss, therefore, assume great significance, for example, in the calculation of groundwater recharge.

Difficulties in understanding the physical nature of the evaporation process, together with ambiguous results from the various types of instrument designed to measure evaporation directly (such as evaporation pans and evaporimeters), led to the development of empirical techniques for estimating evaporation, using generally available climatic data by Thornthwaite in

1948, by Blaney and Criddle in 1950 and by Turc in 1955. These techniques were recognized and acknowledged to give only approximate estimates, but in the absence of simple-to-apply, more theoretically sound methods they provided a useful means of calculating irrigation need and consumptive water use by crops.

Advances in micrometeorology have produced more sophisticated techniques for measuring evaporation. Generally speaking, these are still research techniques requiring far more instrumentation or experimental data than are normally available.

Perhaps the best compromise is the semi-empirical but physically based formulae of Penman developed in 1948, 1952 and subsequently (Penman, 1956; Penman, 1963), or its many derivatives (Monteith, 1965; Thom and Oliver 1977). This embodies the concepts of 'potential transpiration' (PET) from vegetation plentifully supplied with water, and of 'open-water evaporation' (E_O) from an extensive open-water surface. The original formula of Penman developed in 1948 is a combination of the energy balance and aerodynamic methods of measuring evaporation. The energy quantities available for evaporation and for heating the soil-plant-atmosphere system can be equated as follows:

$$R_n = \lambda E + K_a + G \tag{1.4}$$

where R_n = net radiation, λ = product of mass density of water and latent heat of evaporation of water, E = daily lake evaporation, K_a = sensible heat transferred to the air, and G = sensible heat transferred to the soil.

In tropical regions G becomes small in relation to R_n over a day, and may be neglected. R_n can be measured or estimated from incoming solar radiation or from hours of bright sunshine (Grover and McCulloch, 1958), and the problem becomes that of partitioning R_n between sensible heating of the air and the latent heat flux.

The ratio $K_a/(\lambda E)$ is known as the *Bowen ratio* (β). Penman derived an estimate for β by introducing an empirical aerodynamic term E_a and eliminating the need to measure surface temperatures. Evaporation from an open-water surface (E_O) is then given by:

$$E_O = \frac{\Delta H + \gamma E_a}{\Delta + \gamma} \tag{1.5}$$

where

$$E_a = f(u)\,(1 + u_2/160)\,(e_a - e_d), \tag{1.6}$$

and

$$H = (R_n - G)/\lambda \tag{1.7}$$

Here, Δ = the slope of the curve relating saturation vapor pressure to air temperature at mean air temperature,

γ = the psychrometric constant,

f(u) = an empirical constant,

u_2 = mean wind speed at 2 m height above the ground (km/d), and

e_a, e_d = saturation vapor pressure at air temperature and dew point respectively.

On the basis of the Lake Hefner experimental results of the US Navy in 1952, Penman modified the aerodynamic term to:

$$E_a = f(u)\,(0.5 + u_2/160)\,(e_a - e_d) \tag{1.8}$$

justifying the correction on the grounds that the new term gave better agreement with evaporation from a large body of water (Penman, 1956). To estimate PET, Penman first used a reduction factor, which varied seasonally. Averaged over the whole year, a value of 0.75 was derived for western Europe. At a later stage, making use of measurements of the albedo grass, and reinstating the original aerodynamic term to take into account the aerodynamic roughness of short vegetation, a one-step formula was introduced (Penman,1963):

$$PET = \frac{\Delta H + \gamma E_{at}}{\Delta + \gamma} \tag{1.9}$$

where,

$E_{at} = f(u)(1 + u_2/160)(e_a - e_d)$, and

$H = (R_n - G)/\lambda$, with R_n—measured over grass.

This formula has been used to provide an index of evaporation. In practice, it can be expected to give reasonable estimates of ET within the accuracy of the other components of water balance, where water supply to the root zone is not a limiting factor. Simplified calculation methods for the Penman formula can be found in McCulloch (1965), Berry (1964) and Doorembos and Pruitt (1977). Further modifications of the approach can be found in Monteith (1965) and Thom and Oliver (1977).

Where water supply to the root zone is a limiting factor, as in most of the semi-arid tropics, actual evaporation (AE) is considerably less than potential evaporation (PE). The direct methods of measuring AE require complex and expensive equipment, and the indirect methods, such as the water balance of watertight catchments, do not give short-period water use unless soil moisture measurements are available.

An alternative approach has been developed by Bouchet in 1963, and embodies the concept of complementary evaporation. Broadly speaking, this concept states that the difference between AE at a dry site and PE at a wet site, subject to the same radiation input (where water supply is not limiting), is the same as the difference between PE estimated at the dry site (i.e., with the same radiation input but lower humidities and higher air temperatures) and PE calculated for a wet site. This can best be illustrated by a simple diagram, in which annual evaporation is plotted against annual rainfall. The theoretical values of PE are seen to decrease with increasing rainfall at the drier site (location A) until they reach a value PE_0, which is a function of radiation input, temperature and humidity, where soil moisture is not limiting (location B). Variations in the theoretical values of AE are the precise opposite, increasing with increasing rainfall until they reach the same limiting value of PE_0.

Brutsaert and Stricker (1979) used the Priestley-Taylor equation (Priestley and Taylor, 1972) to calculate PE_o. The formula, which was developed to estimate evaporation in the absence of advection, reads:

$$PE_0 = \frac{1.26\Delta R_n}{\Delta + \gamma} \tag{1.10}$$

They used the Penman formula to estimate the local potential evaporation (PE_B). AE can now be found from the Bouchet relationship:

$$AE = 2PE_0 - PE_B \tag{1.11}$$

This approach was tested by Stewart et al (1982), together with an improved Penman formula incorporating a larger aerodynamic term (Thom and Oliver, 1977). They found, using data from 120 tropical stations that the method is only valid when used with climatological data recorded at sites further than 50 km from a coast. At these stations the modified Brutsaert and Stricker formula gave encouraging results, which support the concept of complementary evaporation. Clearly the approach has great potential in areas of sparse data.

1.3.4.1 Potential Evapotranspiration

One aspect of the soil-water budget that involves significant uncertainty and ambiguity is estimating potential evapotranspiration. Just the concept of potential evapotranspiration is ambiguous by itself, as discussed in the next section. Due to limited meteorological data, two simple methods for estimating potential evapotranspiration were considered for the Niger basin study, the Priestley-Taylor and Thornthwaite methods. For the short term simulation (July 1983 to December 1990), a global net radiation data set obtained from NASA facilitated making potential evapotranspiration estimates using the Priestley-Taylor method. For reasons discussed later in this book, the Priestley-Taylor method is considered superior to the Thornthwaite approach; however, it is simpler to apply the Thornthwaite approach to long term average conditions and to selected historical periods because the global net radiation data used in this study are only available from July 1983 to June 1991. It would be nice to have consistent methods for estimating potential evapotranspiration over different time periods so that fair comparisons can be made. Because the Thornthwaite method is more easily applied over different historical time periods, determining whether there are significant differences between predicted runoff using the Priestley-Taylor and Thornthwaite methods is an important question. The conclusion is that there are significant differences and the Priestley-Taylor approach is better. For this study, the average net radiation over the eight-year period when net radiation data were available was taken to be the long term average net radiation. Both Priestley-Taylor and Thornthwaite methods perform poorly in arid regions and the significance of this is briefly discussed.

1.3.4.2 Potential Evaporation vs Potential Evapotranspiration

Thornthwaite in 1948, first used the concept of potential evapotranspiration as a meaningful measure of moisture demand to replace two common surrogates for moisture demand temperature and pan evaporation. Potential evapotranspiration refers to the maximum rate of evapotranspiration from a large area completely and uniformly covered with growing vegetation with an unlimited moisture supply. There is a distinction between the term potential evapo*trans*piration and potential evaporation from a free water surface because factors such as stomatal impedence and plant growth stage influence evapotranspiration but do not influence potential evaporation from free water surfaces.

Brutsaert (1982) notes the remarkable similarity in the literature among observations of water losses from short vegetated surfaces and free water surfaces. He poses a possible explanation that the stomatal impedance to water vapor diffusion in plants may be counterbalanced by larger roughness values. Significant differences have been observed between potential evapotranspiration from tall vegetation and potential evaporation from free water surfaces. The commonly used value of 1.26 in the Priestley-Taylor equation was derived using observations over both open water and saturated land surfaces. For the most part, the term potential evapotranspiration will be used predominantly in this chapter and, as used, includes water loss directly from the soil and/or through plant transpiration.

An additional ambiguity in using the potential evapotranspiration concept is that potential evapotranspiration is often computed based on meteorological data obtained under non-potential conditions. In this study, temperature and net radiation measurements used for calculating potential evapotranspiration in dry areas and for dry periods will be different than the values that would have been observed under potential conditions. The fact that the Thornthwaite and Priestley-Taylor methods have exhibited weak performance at arid sites is related to this ambiguity because the assumptions under which the expressions were derived break down. Poor performance in arid regions is highly relevant to the Niger Basin study in Africa, because of large arid areas in the northern part of the basin. This problem will be addressed a bit further during the detailed discussions of each method.

Although not used directly in this chapter, a brief review of the widely used Penman equation serves as a good starting point for discussing the estimation of potential evapotranspiration.

1.3.4.3 Estimating Actual Evapotranspiration

To estimate the actual evapotranspiration (AET) in the soil-water budget method many investigators have used a *soil-moisture extraction function* or *coefficient of evapotranspiration* (β') which relates the actual rate of evapotranspiration to the potential rate of evapotranspiration based on some function of the current soil moisture content and moisture retention Properties of the soil

$$AET = \beta' \cdot PET \tag{1.12}$$

1.4 THE WATER BALANCE

Water balance or *water budget* or *hydrologic budget* indicates the mass balance in water input and output for a system over a certain time period. The general expression describing the water balance of a watertight catchment over a given period is:

$$P = Q + AET + \Delta S + \Delta G \tag{1.13}$$

where, P is precipitation and Q is the streamflow respectively and can usually be measured directly; AET is actual evapotranspiration; and ΔS, ΔG are changes in soil moisture and groundwater storage respectively.