Hydrology in Practice

Fourth Edition

Elizabeth M. Shaw, Keith J. Beven, Nick A. Chappell and **Rob Lamb**

Spon Press

Hydrology in Practice

Hydrology in Practice is an excellent and very successful introductory text for engineering hydrology students who go on to be practitioners in consultancies, the Environment Agency and elsewhere. It has never been superseded by any other text in this respect.

This fourth edition of *Hydrology in Practice* while retaining all that is excellent about its predecessor, by Elizabeth Shaw, replaces the material on the Flood Studies Report with an equivalent section on the methods of the Flood Estimation Handbook and its revisions. Other completely revised sections on instrumentation and modelling reflect the many changes that have occurred over recent years. The updated text has taken advantage of the extensive practical experience of the staff of JBA Consulting who use the methods described on a day-to-day basis. Topical case studies further enhance the text and the way in which students at undergraduate and MSc level can relate to it. The fourth edition will also have a wider appeal outside the UK by including new material on hydrological processes which also relate to courses in geography and environmental science departments. In this respect the book draws on the expertise of Keith J. Beven and Nick A. Chappell who have extensive experience of field hydrological studies in a variety of different environments, and teaching undergraduate hydrology courses for many years.

Second and final year undergraduates (and MSc) students of hydrology in Engineering, Environmental Science and Geography Departments across the globe, as well as professionals in environmental protection agencies and consultancies, will find this book invaluable. It is likely to be the course text for every undergraduate/MSc hydrology course in the UK and in many cases overseas too.

Elizabeth M. Shaw is now retired but has worked as a lecturer at universities in the UK and Australia, as well as having practised as a hydrologist. **Keith J. Beven** and **Nick A. Chappell** are in the Lancaster Environment Centre at the University of Lancaster, UK. **Rob Lamb** is a consultant with JBA Consulting.

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Fourth edition

Elizabeth M. Shaw, Keith J. Beven,
Nick A. Chappell and Rob Lamb

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Preface to the fourth edition

When the first edition of Elizabeth M. Shaw's *Hydrology in Practice* appeared in 1983, it was immediately perceived to be a valuable addition to the hydrology texts that were then available. It became a standard text in most, if not all, undergraduate and Master's level hydrology courses in the UK and was widely used elsewhere, despite the strong orientation towards the practical training of UK engineering hydrologists. It was Elizabeth's stated intention to ensure that engineers received the training necessary in hydrological measurements and analysis to be able to manage surface and groundwater water resources, floods and droughts in the UK.

Since that time, the organisation of the water industry has changed dramatically in the UK. No longer are there integrated water authorities dealing with supply, waste water, flood risk and water quality. Now we have regulated utility companies with responsibilities for supply and treatment and separate government agencies with responsibilities for water quality standards, licensing of abstractions and effluents, and flood defence. Much less hydrological analysis is done 'in house' by these bodies; much more is commissioned from consultants. The major insurance companies have also taken much more interest in the hydrology of floods and droughts. European Directives, including the Habitats Directive, Water Framework Directive and Floods Directive, have resulted in environmental and sustainability concerns becoming more important relative to purely engineering design issues in water resource management.

Elizabeth has long retired from teaching at Imperial College, but in talking to her in preparing this edition it is clear that she remains concerned about the training of the next generation of engineers and environmental hydrologists. She particularly feels that the recent developments have resulted in a loss of local expertise dealing with local problems using best practice methods. When we received the invitation of Taylor & Francis to prepare a fourth edition of *Hydrology in Practice*, she generously gave us a free hand to change the text and presentation in any way we wished. We have very much intended, however, to keep the applied nature of the text well to the fore, so that it still adhered to the original aims of the first edition. The need is still there, even if the scope of practical training required for hydrologists is now somewhat wider. Some of the methods in the earlier editions have been superceded but we hope that much of the spirit of the original remains.

Indeed, although we have made extensive changes to the text, there is still a lot that is recognisable from the third edition. Thus we have suggested that Elizabeth should remain as the first author for this fourth edition, especially since everyone has simply referred to the earlier editions as 'Shaw' for as long as we can remember.

We feel that this would be a suitable recognition of her valuable contribution to the training of several generations of hydrology students. We have therefore maintained the structure of the previous editions in grouping the chapters into Parts, concerned with Hydrological Measurements (Chapters 2–8), Hydrological Analysis (Chapters 9–15) and Engineering Applications (Chapters 16–19). We have tried to look to the future in adding an additional chapter on The Future of Hydrology in Practice.

Special thanks are due to Jeremy Benn and the staff of JBA Consulting. This is a consultancy that has grown with the reorganisation of the water industry in the UK and that has been at the forefront in developing and applying new methods of hydrological and hydraulic analysis. This edition has greatly benefited from the experience of practical applications within the company and we would particularly like to recognise the contributions of David Archer, Jeremy Benn, Eleanor Charles, Mandy Crossley, John Dudley, Paul Dunning, Paul Ecclestone, Duncan Faulkner, Barry Hankin, Peter Henrys, Neil Hunter, Caroline Keef, John Mawdsley, Steve Rose, Zdenka Rosolova, Vicky Shackle, Judith Stunell, Sebastien Tellier, Simon Waller, Paul Wass and Maxine Zaidman.

We are also grateful to the following individuals for their assistance: Andy Binley and Hao Zhang of Lancaster University, Paul Quinn of Newcastle University, Tracey Haxton, and Andy Young of Wallingford HydroSolutions, Janet Bromley of United Utilities, and Jamie Hannaford, Alison Kay, Gwyn Rees, Nick Reynard of CEH Wallingford.

We are grateful to many organisations and individuals for permission to use their line drawings and photographs. Each image is individually acknowledged within the associated figure caption. While we have illustrated the text on hydrological measurements with representative equipment that is currently available in the UK, it should be noted that equipment of similar or alternative design will be available from other companies and no specific endorsement by the authors is implied.

Keith J. Beven
Nick A. Chappell
Rob Lamb

Chapter 1

The hydrological cycle and hydrometeorology

The history of the evolution of hydrology as a multi-disciplinary subject, dealing with the occurrence, circulation and distribution of the waters of the Earth, has been presented by Biswas (1970). Man's need for water to sustain life and grow food crops was well appreciated throughout the world wherever early civilization developed. Detailed knowledge of the water management practices of the Sumarians and Egyptians in the Middle East, of the Chinese along the banks of the Hwang-Ho and of the Aztecs in South America continues to grow as archaeologists uncover and interpret the artefacts of such centres of cultural development. It was the Greek philosophers who were the first serious students of hydrology, and thereafter, scholars continued to advance the understanding of the separate phases of water in the natural environment. However, it was not until the seventeenth century that the work of the Frenchman, Perrault, provided convincing evidence of the form of the hydrological cycle which is currently accepted: measurements of rainfall and river flow in the catchment of the upper Seine published in 1694 (Dooge, 1959) proved that quantities of rainfall were sufficient to sustain river flow.

Hydrology as an academic subject became established within institutions of higher education in the 1940s. Valuable research contributions to the subject had been reported earlier but the expansion in the more widespread applications of hydrology resulted in at least five textbooks being published in that decade in the United States.

Over the last 50 years, advances in sensor technology coupled with the development of numerical models representing hydrological processes have led to a reappraisal of the content and definition of hydrology. Today's scientific hydrologists and engineering hydrologists now appreciate the need to combine accurate field measurement with appropriate numerical models. Equally, there is an awareness of the controlling influence of hydrometeorology on the water pathways that comprise the hydrological cycle at catchment and global scales.

1.1 The hydrological cycle and water pathways

The driving force of the natural circulation of water is derived from the radiant energy received from the Sun. The bulk of the Earth's water is stored on the surface in the oceans (Table 1.1) and hence it is logical to consider the hydrological cycle as beginning with the direct effect of the Sun's radiation on this largest reservoir. Heating of the sea surface causes *evaporation*, the transfer of water from the liquid to the gaseous

Table 1.1 One estimate of global water distribution

Store	*Volume (1000 km^3)*	*Per cent of total water*	*Per cent of fresh water*
Oceans, seas and bays	1 338 000	96.5	–
Ice caps, glaciers and permanent snow	24 064	1.74	68.7
Groundwater	23 400	1.7	–
Fresh	(10 530)	(0.76)	30.1
Saline	(12 870)	(0.94)	–
Soil moisture	16.5	0.001	0.05
Ground ice and permafrost	300	0.022	0.86
Lakes	176.4	0.013	–
Fresh	(91.0)	(0.007)	0.26
Saline	(85.4)	(0.006)	–
Atmosphere	12.9	0.001	0.04
Swamp water	11.47	0.0008	0.03
Rivers	2.12	0.0002	0.006
Biological water	1.12	0.0001	0.003

Source: Gleick, P. H. (1996) Water resources. In *Encyclopedia of Climate and Weather*, ed. by S. H. Schneider, Oxford University Press, New York, vol. 2, pp. 817–823.

state, to form part of the atmosphere. It remains mainly unseen in atmospheric storage for an average of 10 days. Through a combination of circumstances, the water vapour changes back to the liquid state again through the process of *condensation* to form clouds and, with favourable atmospheric conditions, *precipitation* (rain, snow, etc.) is produced either to return directly to the ocean storage or to embark on a more devious route to the oceans via the land surface. Snow may accumulate in polar regions or on high mountains and consolidate into ice, in which state water may be stored naturally for very long periods. In more temperate lands, rainfall may be intercepted by vegetation from which some of the intercepted water may return at once to the air by wet-canopy evaporation. A significant proportion of the rainfall that reaches the land surface will return to the atmosphere by transpiration via plants, while the remainder travels over or beneath the land surface towards rivers by mechanisms described as runoff[1] generation pathways.

1.2 Pathways generating river flow

In the later nineteenth and early twentieth century, it began to be recognised that different parts of a catchment area might produce different amounts of river flow. Again this was perhaps expressed first in France in the work of Imbeaux in a study of the Durance basin published in 1892. He tried to take account of the role of river flow generation at different distances from a catchment outlet in controlling the shape of the hydrograph, and of elevation in controlling the patterns of snowmelt during the melt season. The idea of delay in runoff reaching the catchment outlet can be represented in terms of a *time–area histogram*. It was later developed into the first storm event rainfall to river flow model to be widely used around the world that we now know as the *unit hydrograph* (see Section 12.5).

Use of the unit hydrograph for practical applications, however, requires that we try to estimate the proportion of storm rainfall that contributes to the *storm hydrograph* for a particular event (i.e. that river flow which appears soon after rainfall). It has long been known that not all the rainfall falling in an event contributes to the storm hydrograph. Some contributes to a much slower subsurface pathway or is lost back to the atmosphere by evapo-transpiration. Thus application of the unit hydrograph concepts required a rather arbitrary separation of the hydrograph into so-called *stormflow* and *baseflow*. This led to a rather easy assumption that the storm runoff in a river was made up of rainfall from the particular rain-event or snowmelt water.

In fact we now know that the situation is somewhat more complicated than that easy assumption because analyses of environmental tracers since the 1970s have shown that, in many environments, not all the storm hydrograph is made up of rainfall that fell in that storm (see Section 11.3). Some of the hydrograph comes from water that was already stored in the catchment prior to the rainfall event. That water is displaced from storage into the stream channels during the event as a result of rainfall infiltrating into the subsurface. This is really one of the most important conceptual advances in scientific hydrology since it has very important implications for understanding *hillslope hydrology*, water quality variations and ecological impacts of storm events.

The history of river flow generation concepts often starts with the ideas of Robert Elmer Horton, probably the most influential American hydrologist of the twentieth century. In a paper published in 1933 (and reproduced in Beven, 2006) Horton first expressed a concept of hillslope hydrology based on the idea that the storm hydrograph is made up of rainfall in excess of the infiltration capacity of the soil. This is the concept of *infiltration excess (or Hortonian) overland flow*. This idea leads to a nice simple interpretation of catchment response. If we know the volume of rainfall in an event, and we know the volume of *stormflow* (here meaning that river flow proportion above the hydrograph separation line; Fig. 12.5, Section 12.3) recorded at a river gauging site, then the difference must be what was infiltrated (on average) into the soil. This allows the infiltration capacity of the soil to be back-calculated (subject to some simple assumptions of how it might vary over time). This information can then be used in a simple runoff model to predict what might happen under different conditions. Combining this prediction of how much runoff will be generated by a given rainfall, and the unit hydrograph to predict the timing of the runoff is a technique that is still used to the present day (it still underlies some aspects of the UK Flood Estimation Handbook, for example, see Chapter 16).

The problem is, of course, that it is wrong. Some catchments in arid areas under high rainfall intensities, or the extreme case of impermeable surfaces in urban areas, might work like this, though even then it is unlikely that infiltration excess overland flow will occur everywhere, and overland flow generated on one part of a slope might later infiltrate further downslope. It is even unlikely that Horton saw this type of infiltration-excess overland flow in his own experimental catchment in New York State (Beven, 2004). However, as we have already noted, in very many catchments, much of the storm hydrograph is made up of displaced pre-event storage. Thus, other concepts of river flow generation are needed.

The first real reconsideration of the Hortonian concept was by Roger P. Betson in 1964. Betson worked for the Tennessee Valley Authority in the United States and realised that in the forested catchments of the Appalachians, there was no way that

infiltration excess overland flow could occur everywhere, except perhaps in the most extreme rainfall events. He therefore suggested that overland flow would be generated on only part of the hillslopes and that since infiltration rates of soils tend to be lower when the soil is wetter (Section 5.1.1) and, as a result of downslope flows between events, soils will tend to be wetter in the lower parts of hillslopes, then the runoff generation would be most likely at the bottom of hillslopes close to the stream channels. He inferred from the analyses of storm runoff volumes that the proportion of the catchment generating overland flow could be quite small (as low as 2–4 per cent) in some catchments.

At about the same time, John Hewlett, working at the Coweeta catchments in North Carolina, suggested that the infiltration capacities of the soils in that area were so large that it was extremely unlikely that any runoff would be seen over the surface of the soil. Yet storm hydrographs were still recorded. He suggested that the storm runoff therefore must be generated by subsurface flows and by rainfall directly on to the stream channel and immediate riverside area. He also suggested that the water contributing to the river was not necessarily the rainfall, invoking a concept of so-called 'translatory flow' to explain the displacement of stored water by the infiltration of the rainwater. In fact, a previous Director of the Coweeta Laboratory, Charles Hursh, had expressed much the same idea in the 1930s, and had coined the term *subsurface stormflow* for this type of river flow generation mechanism. The concept of translatory flow had also already been mentioned in the 1930s but these concepts had been dominated by the Hortonian paradigm in the later engineering literature.

There are some circumstances, however, when overland flow can be generated on soils with high infiltration capacities. This is when the soil becomes saturated by a combination of downslope flow within a hillslope and rain falling on saturated areas. In fact, downslope flows can maintain the lower parts of hillslopes at, or close to, saturation for long periods of time in some circumstances, so that only small amounts of rainwater might be required before overland flow is generated. This will particularly be the case in relatively shallow soils overlying an impermeable base, and where there are convergent flow lines into the hillslope hollows. This was first demonstrated by the work of Tom Dunne in the Sleepers River catchments in Vermont in the late 1960s. He showed how saturated areas in the catchment could persist for long periods of time, how they varied seasonally, being most extensive at the end of the snowmelt season in Vermont, and how they were largely controlled by the patterns of downslope flow on the hillslopes. This type of overland flow became known as *saturation overland flow* that was generated on a *variable contributing area* in the catchment. In some cases the resulting overland flow will also have a component of *return flow*, which subsurface water forced back on to the surface through a seepage face; and one of the earliest studies of environmental tracers in storm runoff, by Mike Sklash and Bob Farvolden (1979), provided evidence to reinforce this concept. In an area of river flow generation in one of their study catchments, they showed that tracer concentrations were sometimes indicative of a rainfall source and at other times indicative of a subsurface, pre-event storage source.

Work elsewhere has revealed further complications. Darrell Weyman (1970) working in the East Twin catchment in the Mendips, UK, showed that saturated contributing areas could arise without the soil being completely saturated but where saturation

built up above a soil horizon of lower permeability. He also showed that subsurface contributions to the stream channel could be hugely variable in space and might be associated with zones of higher soil permeability within a very heterogeneous soil. Bunting (1961) had earlier called such preferential pathways *percolines* and he treated them as a subsurface extension of the dendritic (tree-like) channel network. Recent modelling work has shown how, in heterogeneous soils, subsurface flow might be simulated as being channelled into channel network-like structures (e.g. Weiler and McDonnell, 2004), while tracer work has shown that, in some catchments, fast responding subsurface pipes produce storm runoff that may be made up predominantly of pre-event stored water rather than rainwater (Sklash *et al.*, 1996).

It is perhaps worth finishing this brief summary of river flow generation pathways by saying that the different major concepts shown in Fig. 1.1 are not mutually exclusive. They might all occur in different events in the same catchment, or in the same event in different parts of a catchment, depending on the rainfall intensities; prior wetness of the catchment (*antecedent conditions*); topography of the hillslopes; type, structure and heterogeneity of the soil, regolith and rock; existence of percolines; channel

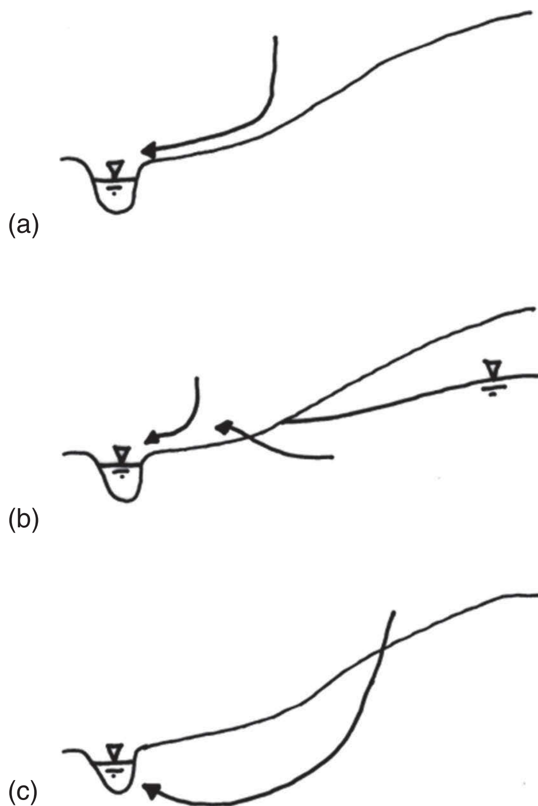


Fig. 1.1 River flow generation pathways for systems dominated by (a) infiltration-excess overland flow, (b) saturation overland flow, and (c) subsurface flow.

density; and other factors. An excellent review of where and when different types of runoff production might occur is given by Tom Dunne (1978) and Wilfred Brutseart (2005).

There is an underlying research question about how, over long periods of time, these different factors might be linked to the long-term development of the catchment soils, topography and vegetation cover, and how, in recent times, people might have affected the nature of the river flow generation processes through land management practices and urbanisation. Such questions are not yet fully resolved, and it is perhaps unlikely that they will ever be properly resolved given the complexities of short-term and long-term changes to which catchments have been subjected in different environments.

Fortunately, this is not a barrier to hydrological analysis and prediction. In many cases we are only interested in predicting river flow, and do not need to worry too much about the water pathways. This is one reason why unit hydrograph concepts have survived so long: if we can match river flow volumes and timings using these simple concepts, then we may be able to make some useful predictions even if the details of the pathways are incorrect. There are situations, however, particularly in understanding water quality variations, where it may be critical to appreciate the different surface and subsurface pathways. In such cases, an appreciation for the different mechanisms of river flow generation described above will be important. We will return to this in the discussion of predictive rainfall-runoff models in Chapter 12. First, we need to give an overview of how hydrometeorology regulates these river flow generation pathways and the pathways of evapo-transpiration (Chapter 10).

1.3 Hydrometeorological control of hydrological pathways

The science of meteorology has long been recognised as a separate discipline, though students of the subject usually come to it from a rigorous training in physics or mathematics. The study of *hydrometeorology* may be seen as a branch of hydrology linking the fundamental knowledge of the meteorologist with the needs of the hydrologist. In this text, hydrometeorology is taken to be the study of precipitation and evaporation, the two fundamental phases in the hydrological cycle, which involve processes in the atmosphere, and at the Earth's surface/atmosphere interface.

The hydrologist will usually be able to call upon the services of a professional meteorologist for weather forecasts and for special studies, e.g. the magnitude of extreme rainfalls. However, a general understanding of precipitation and evaporation is essential if the hydrologist is to appreciate the complexities of the atmosphere and the difficulties that the meteorologist often has in providing answers to questions of quantities and timing. A description of the properties of the atmosphere and of the main features of solar radiation will provide the bases for considering the physics of evaporation and the formation of precipitation.

1.3.1 The atmosphere

The atmosphere forms a distinctive protective layer about 100 km thick around the Earth. Although both air pressure and density decrease rapidly and continuously with

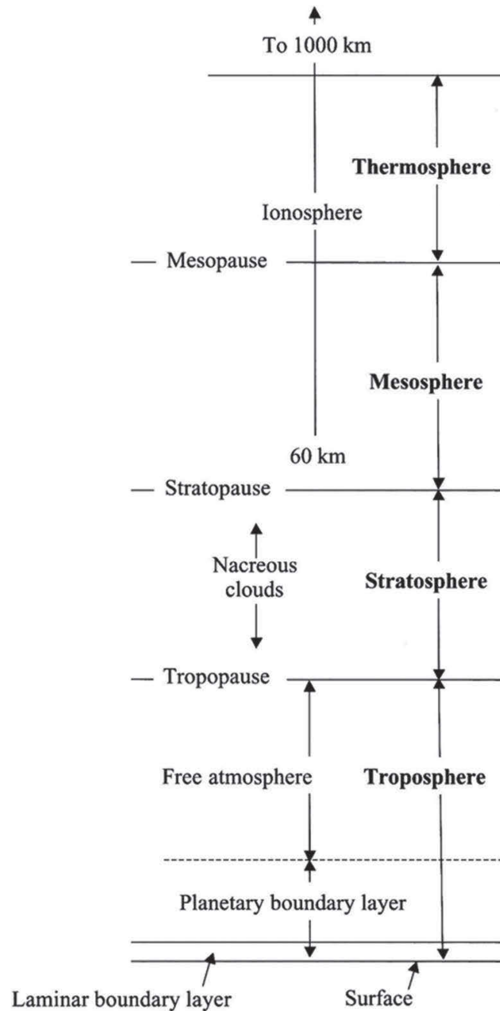


Fig. 1.2 Structure of the atmosphere. (Adapted from Strangeways, I. (2007) *Precipitation: Theory, Measurement and Distribution*, Cambridge University Press, Cambridge.)

increasing altitude, the temperature varies in an irregular but characteristic way. The layers of the atmosphere, 'spheres', are defined by this temperature profile. After a general decrease in temperature through the *troposphere* (Fig. 1.2), the rise in temperature from heights of 20–50 km is caused by a layer of ozone, which absorbs short-wave solar radiation, releasing some of the energy as heat.

To the hydrologist, the troposphere is the most important layer because it contains 75 per cent of the weight of the atmosphere and virtually all its moisture. The meteorologist, however, is becoming increasingly interested in the stratosphere and mesosphere, since it is in these outer regions that some of the disturbances affecting the troposphere and the Earth's surface have their origins.

The height of the *tropopause*, the boundary zone between the troposphere and the stratosphere, is at about 11 km, but this is an average figure, which ranges from about 8 km at the Poles to about 16 km at the Equator. Seasonal variations also are caused by changes in pressure and air temperature in the atmosphere. In general, when surface temperatures are high and there is a high sea-level pressure, then there is a tendency for the tropopause to be at a high level. On average, the temperature from ground level to the tropopause falls steadily with increasing altitude at the rate of $6.5^\circ C\,km^{-1}$. This is known as the *lapse rate*. Some of the more hydrologically pertinent characteristics of the atmosphere as a whole are now defined in more precise terms.

1.3.1.1 Atmospheric pressure and density

The meteorologist's definition of atmospheric pressure is 'the weight of a column of air of unit area of cross-section from the level of measurement to the top of the atmosphere'. More specifically, pressure may be considered to be the downward force on a unit horizontal area resulting from the action of gravity (g) on the mass (m) of air vertically above.

At sea level, the average atmospheric pressure (p) is 100 kPa (1 bar or 100 000 $N\,m^{-2}$). A pressure of 100 kPa is equivalent to 760 mm of mercury; the average reading on a standard mercury barometer. Measurements of atmospheric pressure are usually given in millibars (mb). It is common meteorological practice to refer to heights in the atmosphere by their average pressure in millibars, e.g. the top of the stratosphere (the stratopause) is at the 1 mb level. The air density (ρ) may be obtained from the expression $\rho = p/RT$, where R is the specific gas constant for dry air ($0.29\,kJ\,kg^{-1}\,K^{-1}$) and $T\,K$ is the air temperature. At sea level, the average $T = 288$ K and thus $\rho = 1.2\,kg\,m^{-3}$ (or $1.2 \times 10^{-3}\,g\,cm^{-3}$) on average at sea level. Air density falls off rapidly with height. Unenclosed air, a compressible fluid, can expand freely, and as pressure and density decrease with height indefinitely, the limit of the atmosphere becomes indeterminate. Within the troposphere however, the lower pressure limit is about 100 mb. At sea level, pressure variations range from about 940 to 1050 mb; the average sea level pressure around the British Isles is 1013 mb. Pressure records form the basis of the meteorologist's synoptic charts with the patterns formed by the *isobars* (lines of equal pressure) defining areas of high and low pressure (anticyclones and depressions, respectively). Interpretation of the charts plotted from observations made at successive specified times enables the changes in weather systems to be identified and to be forecast ahead. In addition to the sea level measurements, upper air data are plotted and analysed for different levels in the atmosphere.

1.3.1.2 Chemical composition

Dry air has a very consistent chemical composition throughout the atmosphere up to the mesopause at 80 km. The proportions of the major constituents are as shown in Table 1.2. The last category contains small proportions of other inert gases and, of particular importance, the stratospheric layer of ozone which filters the Sun's radiation. Small quantities of hydrocarbons, ammonia and nitrates may also exist temporarily in the atmosphere. Man-made gaseous and particulate pollutants are found particularly in areas of heavy industry, and can have considerable effects on local

Table 1.2 Major constituents of air

	Percentage (by mass)
Nitrogen	75.51
Oxygen	23.15
Argon	1.28
Carbon dioxide etc	0.06

weather conditions. Traces of radioactive isotopes from nuclear fission also contaminate the atmosphere. Although there is no evidence that isotopes have a significant effect on weather, their presence has been found useful in tracing the movement of water through the hydrological cycle.

1.3.1.3 Water vapour

The amount of water vapour in the atmosphere (Table 1.3) is directly related to the temperature and thus, although lighter than air, water vapour is restricted to the lower layers of the troposphere because temperature decreases with altitude. The distribution of water vapour also varies over the Earth's surface according to temperature, and is lowest at the Poles and highest in equatorial regions. The water vapour content or *humidity* of air is usually measured as a vapour pressure, and the units used are millibars (mb).

Several well-recognised physical properties concerned with water in the atmosphere are defined to assist understanding of the complex changes that occur in the meteorological phases of the hydrological cycle.

(a) *Saturation.* Air is said to be saturated when it contains the maximum amount of water vapour it can hold at its prevailing temperature. The relationship between saturation vapour pressure (e) and air temperature is shown in Fig. 1.3. At typical temperatures near the ground, e ranges from 5 to 50 mb. At any temperature $T = T_a$, saturation occurs at corresponding vapour pressure $e = e_a$.

Table 1.3 Average water vapour values for latitudes with temperate climates (volume %)

Height (km)	*Water vapour*
0	1.3
1	1.0
2	0.69
3	0.49
4	0.37
5	0.27
6	0.15
7	0.09
8	0.05

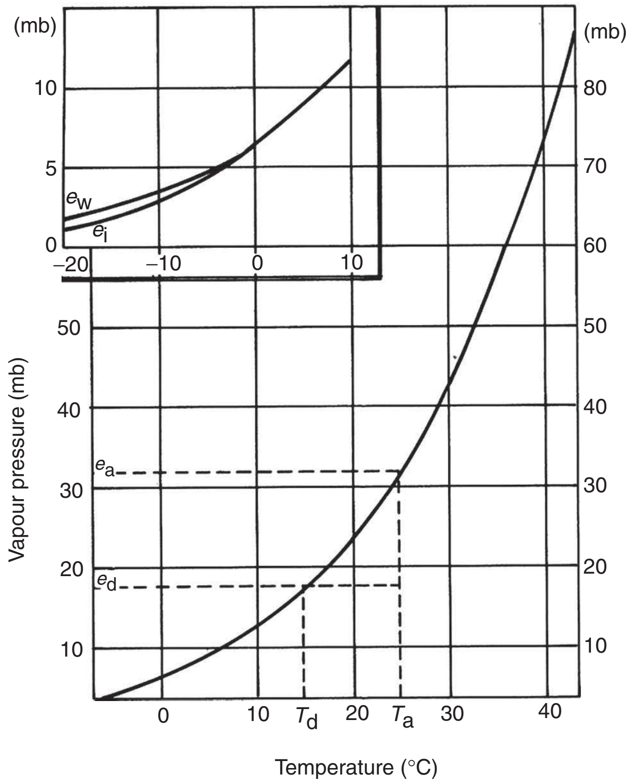


Fig. 1.3 Saturation pressure and air temperature, where $e_a - e_d$ is the saturation deficit and T_d is the dew point temperature.

Meteorologists acknowledge that saturated air may take up even more water vapour and become *supersaturated* if it is in contact with liquid water in a sufficiently finely divided state (e.g. very small water droplets in clouds). At sub-zero temperatures, there are two saturation vapour pressure curves, one with respect to water (e_w) and one with respect to ice (e_i; Fig. 1.3, inset). In the zone between the curves, the air is unsaturated with respect to water but supersaturated with respect to ice. This is a common condition in the atmosphere as will be seen later.

(b) *Dew point* is the temperature, T_d, at which a mass of unsaturated air becomes saturated when cooled, with the pressure remaining constant. In Fig. 1.3, if the air at temperature T_a is cooled to T_d, the corresponding saturation vapour pressure, e_d, represents the amount of water vapour in the air.

(c) *Saturation deficit* is the difference between the saturation vapour pressure at air temperature, T_a, and the actual vapour pressure represented by the saturation vapour pressure at T_d, the dew point. The saturation deficit, $e_a - e_d$, represents the further amount of water vapour that the air can hold at the temperature, T_a, before becoming saturated.

(d) *Relative humidity* is the relative measure of the amount of moisture in the air to the amount needed to saturate the air at the same temperature, i.e. e_d/e_a, represented as a percentage. Thus, if $T_a = 30°C$ and $T_d = 20°C$, relative humidity is

$$\frac{e_d}{e_a} \times 100 = \frac{23\,\text{mb}}{42.5\,\text{mb}} \times 100 = 54 \text{ per cent}$$

(e) *Absolute humidity* (ρ_w) is generally expressed as the mass of water vapour per unit volume of air at a given temperature and is equivalent to the water vapour density. Thus, if a volume $V\,\text{m}^3$ of air contains m_{wv} g of water vapour,

$$\rho_w = \frac{\text{Mass of water vapour (g)}}{\text{Volume of air}\,(\text{m}^3)} = \frac{m_{wv}}{V}(\text{g}\,\text{m}^{-3})$$

(f) *Specific humidity* (SH) relates the mass of water vapour (m_{wv} g) to the mass of moist air (in kg) in a given volume; this is the same as relating the absolute humidity ($\text{g}\,\text{m}^{-3}$) to the density of the same volume of unsaturated air ($\rho\,\text{kg}\,\text{m}^{-3}$):

$$SH = \frac{m_{wv}(\text{g})}{(m_{wv} + m_d)(\text{kg})} = \frac{\rho_w}{\rho}(\text{g}\,\text{kg}^{-1})$$

where m_d is the mass (kg) of the dry air.

(g) *Precipitable water* is the total amount of water vapour in a column of air expressed as the depth of liquid water in millimetres over the base area of the column. Assessing this amount is a specialised task for the meteorologist. The precipitable water gives an estimate of maximum possible rainfall, though has the unreal assumption of total condensation and neglects the effect of advection.

In a column of unit cross-sectional area, a small thickness, dz, of moist air contains a mass of water given by:

$$dm_{wv} = \rho_w\,dz$$

Thus, in a column of air from heights $z1$ to $z2$, corresponding to pressures $p1$ and $p2$: the total mass of water m_w is

$$\int_{z_1}^{z_2} \rho_w\,dz$$

Also, $dp = -\rho g dz$ and, by rearrangement, $dz = -dp/\rho g$. Thus:

$$m_w = -\int_{p_1}^{p_2} \frac{\rho_w}{\rho_g}\,dp$$
$$= \frac{1}{g}\int_{p_1}^{p_2} q dp$$

Allowing for the conversion of the mass of water (m_w) to equivalent depth over a unit cross-sectional area, the precipitable water is given by:

$$W\,(\text{mm}) = \frac{0.1}{g}\int_{p_1}^{p_2} SHdp$$

where p is in mb, SH in g kg^{-1} and $g = 9.81$ m s^{-2}.

In practice, the integration cannot be performed since q is not known as a function of p. A value of W is obtained by *summing* the contributions for a sequence of layers in the troposphere from a series of measurements of the specific humidity $\bar{q}$ at different heights and using the average specific humidity $\bar{q}$ over each layer with the appropriate pressure difference:

$$W\,(\text{mm}) = \frac{0.1}{g}\sum_{p_1}^{p_2} \overline{SH}\Delta p$$

Example. From a radiosonde (balloon) ascent, the pairs of measurements of pressure and specific humidity shown in Table 1.4 were obtained. The precipitable water in a column of air up to the 250 mb level is calculated ($g = 9.81$ m s^{-2}).

1.3.2 Solar radiation

The main source of energy at the Earth's surface is radiant energy from the Sun, termed solar radiation or insolation. It is the solar radiation impinging on the Earth that fuels the heat engine driving the hydrological cycle. The amount of radiant energy received at any point on the Earth's surface (assuming no atmosphere) is governed by the following well-defined factors.

(a) *The solar output.* The Sun, a globe of incandescent matter, has a gaseous outer layer about 320 km thick and transmits light and other radiations towards the Earth from a distance of 145 million km. The rate of emission of energy is shown in Fig. 1.4 but only a small fraction of this is intercepted by the Earth. Half the total energy emitted by the Sun is in the visible light range, with wavelengths from 0.4 to 0.7 μm. The rest arrives as ultraviolet or infrared waves, from 0.25 up to 3.0 μm.

The maximum rate of the Sun's emission (10 500 kW m^{-2}) occurs at 0.5 μm wavelength in the visible light range. Although there are changes in the solar output

Table 1.4

Pressure (mb)	1005	850	750	700	620	600	500	400	250
Specific humidity q (g kg^{-1})	14.2	12.4	9.5	7.0	6.3	5.6	3.8	1.7	0.2
$P_n - P_{n+1} = \Delta p$		155	100	50	80	20	100	100	150
Mean SH $= \overline{SH}$		13.30	10.95	8.25	6.65	5.95	4.70	2.75	0.95
$\overline{SH}\Delta p$		2061.5	1095.0	412.5	532.0	119.0	470.0	275.0	142.5

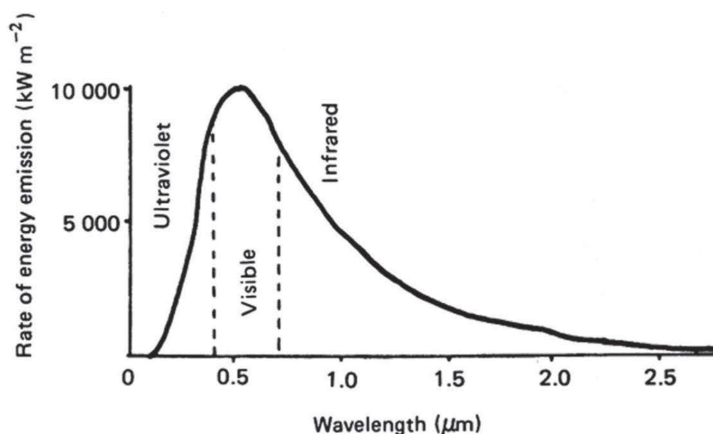


Fig. 1.4 Solar radiation.

associated with the occurrence of sunspots and solar flares, these are disregarded in assessing the amount of energy received by the Earth. The total solar radiation received in unit time on unit area of a surface placed at right angles to the Sun's rays at the Earth's mean distance from the Sun is known as the *solar constant*. The average value of the solar constant is $1.39\,kW\,m^{-2}$ ($1.99\,cal\,cm^{-2}\,min^{-2}$).

(b) *Distance from the Sun.* The distance of any point on the Earth's surface from the Sun is changing continuously owing to the Earth's eccentric orbit. The Earth is nearest the Sun in January at perihelion and furthest from it in July at aphelion. The solar constant varies accordingly.

(c) *Altitude of the Sun.* The Sun's altitude above the horizon has a marked influence on the rate of solar radiation received at any point on the Earth. The factors determining the Sun's altitude are latitude, season and time of day.

(d) *Length of day.* The total amount of radiation falling on a point of the Earth's surface is governed by the length of the day, which itself depends on latitude and season.

1.3.2.1 Atmospheric effects on solar radiation

The atmosphere has a marked effect on the energy balance at the surface of the Earth. In one respect it acts as a shield protecting the Earth from extreme external influences, but it also prevents immediate direct loss of heat. Thus it operates as an energy filter in both directions. The interchanges of heat between the incoming solar radiation and the Earth's surface are many and complex. There is a loss of energy from the solar radiation as it passes through the atmosphere known as *attenuation*. Attenuation is brought about in three principal ways as follows.

(a) *Scattering.* About 9 per cent of incoming radiation is scattered back into space through collisions with molecules of air or water vapour. A further 16 per cent are also scattered, but reach the Earth as diffuse radiation, especially in the shorter wavelengths, giving the sky a blue appearance.

(b) *Absorption.* Fifteen per cent of solar radiation is absorbed by the gases of the atmosphere, particularly by the ozone, water vapour and carbon dioxide. These gases absorb wavelengths of less than 0.3 μm only, and so very little of this radiation penetrates below an altitude of 40 km.

(c) *Reflection.* On average, 33 per cent of solar radiation is reflected from clouds and the ground back into space. The amount depends on the *albedo* (α) of the reflecting surfaces. White clouds and fresh white snow reflect about 90 per cent of the radiation ($r = 0.9$), but a dark tropical ocean under a high sun absorbs nearly all of it ($\alpha \to 0$). Between these two extremes is a range of surface conditions depending on roughness, soil type and water content of the soil. The albedo of the water surface of a reservoir is usually assumed to be 0.05, and of a short grass surface, 0.25.

1.3.2.2 Net radiation

As a result of the various atmospheric losses, only about 43 per cent of solar (short-wave) radiation reaches the Earth's surface, where most is absorbed and heats the land and oceans. The Earth itself radiates energy in the long-wave range (Fig. 1.5) and this long-wave radiation is readily absorbed by the atmosphere. The Earth's surface emits more than twice as much energy in the infrared range as it receives in short-wave solar radiation.

The balance between incoming and outgoing radiation varies from the Poles to the Equator. There is a net heat gain in equatorial regions and a net heat loss in polar regions. Hence, heat energy travels through circulation of the atmosphere from lower to higher latitudes. Further variations occur because the distribution of the continents and oceans leads to differential heating of land and water.

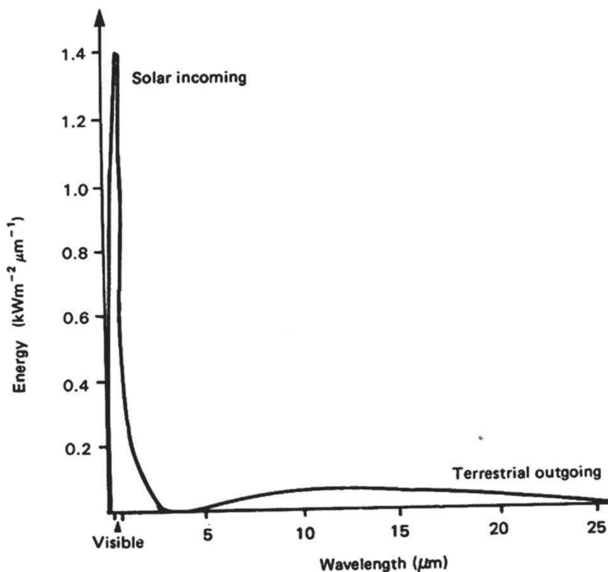


Fig. 1.5 Solar and terrestrial radiation.

Table 1.5 Average radiation values for selected latitudes ($W\,m^{-2}$)

	July season			*January season*		
	R_a	R_o	R_N	R_a	R_o	R_N
50° N	250	210	40	70	190	−120
Equator	280	240	40	310	240	70
30° S	170	220	−50	320	230	90

The amount of energy available at any particular point on the Earth's surface for heating the ground and lower air layers, and for the evaporation of water, is called the *net radiation*.

The net radiation R_N may be defined by the equation:

$$R_N = R_a - \alpha(R_a) + R_l - R_o$$

where R_a is the incoming short-wave (solar) radiation, α is the albedo, R_l is the incoming long-wave radiation and R_o is the outgoing long-wave radiation.

Incoming long-wave radiation comes from clouds (from absorbed solar radiation), and this has the following effects in the net radiation equations. In clear conditions, $R_l \approx (0.6 \text{ to } 0.8)R_o$, thus $R_l - R_o$ gives a net loss of long-wave radiation. For cloudy conditions, $R_l \approx R_o$ and $R_l - R_o$ becomes 0.

More significant are diurnal variations in net radiation, which is the primary energy source for evaporation. At night, $S = 0$ and R_l is smaller or negligible so that $R_N \approx R_o$. In other words, net radiation is negative, and there is a marked heat loss, which is particularly noticeable when the sky is clear.

Some average values of solar (R_a), terrestrial (R_o) and net (R_N) radiation for points on the earth's surface are given in Table 1.5.

1.4 Evaporation

Evaporation is the primary process of water transfer in the hydrological cycle. The oceans contain 95 per cent of the Earth's water and constitute a vast reservoir that remains comparatively undisturbed. From the surface of the seas and oceans, water is evaporated and transferred to temporary storage in the atmosphere, the first stage in the hydrological cycle.

1.4.1 Factors affecting evaporation

To convert liquid water into gaseous water vapour at the same temperature a supply of energy is required (in addition to that possibly needed to raise the liquid water to that temperature). The *latent heat of vaporization* ($2.6 \times 10^6\,J\,kg^{-1}$) must be added to the liquid molecules to bring about the change of state. The energy available for evaporation is the net radiation obtaining at the water surface and is governed by local conditions of solar and terrestrial radiation.

The rate of evaporation is dependent on the temperature at the evaporating surface and that of the ambient air. It also depends on the vapour pressure of the existing water

vapour in the air, since this determines the amount of additional water vapour that the air can absorb. From the saturation vapour pressure and air temperature relationship shown in Fig. 1.3, it is clear that the rate of evaporation is dependent on the saturation deficit. If the water surface temperature, T_s, is equal to the air temperature, T_a, then the saturated vapour pressure at the surface, e_s, is equal to e_a. The saturation deficit of the air is given by $(e_s - e_d)$, where e_d is the measure of the actual vapour pressure of the air at T_a.

As evaporation proceeds, the air above the water gradually becomes saturated and, when it is unable to take up any more moisture, evaporation ceases. The replacement of saturated air by drier air would enable evaporation to continue. Thus, wind speed is an important factor in controlling the rate of evaporation. The roughness of the evaporating surface is a subsidiary factor in controlling the evaporation rate because it affects the turbulence of the air flow.

In summary, evaporation from an open water surface is a function of available energy, the net radiation, the temperatures of surface and air, the saturation deficit and the wind speed. The evaporation from a vegetated surface is a function of the same meteorological variables, but it is also dependent on the presence of *negative pressure potential* (Section 6.2) within the soil or regolith. From a land surface, it is a combination of the evaporation of liquid water from precipitation collected on the land surface, from wetted vegetation surfaces and the transpiration of water by plants. Methods for the measurement of evaporation quantities are presented in detail in Chapter 4, and methods of analysis in Chapter 10.

1.5 Precipitation

The moisture in the atmosphere, although forming one of the smallest storages of the Earth's water, is the most vital source of fresh water for mankind. Water is present in the air in its gaseous, liquid and solid states as water vapour, cloud droplets and ice crystals, respectively.

The formation of precipitation from the water as it exists in the air is a complex and delicately balanced process. If the air was pure, condensation of the water vapour to form liquid water droplets would occur only when the air became greatly supersaturated. However, the presence of small airborne particles called *aerosols* provides nuclei around which water vapour in normal saturated air can condense. Many experiments, both in the laboratory and in the open air, have been carried out to investigate the requisite conditions for the change of state. Aitken (Mason, 1975) distinguished two main types of condensation nuclei: *hygroscopic particles* having an affinity for water vapour, on which condensation begins before the air becomes saturated (mainly salt particles from the oceans); and *non-hygroscopic particles* needing some degree of supersaturation, depending on their size, before attracting condensation. This latter group derives from natural dust and grit from land surfaces and from man-made smoke, soot and ash particles.

Condensation nuclei range in size from a radius 10^{-3} μm for small ions to 10 μm for large salt particles. The concentration of aerosols in time and space varies considerably. A typical number for the smallest particles is 40 000 per cm^3, whereas for giant nuclei of more than 1 μm radius there might be only 1 per cm^3. Large hygroscopic salt nuclei are normally confined to maritime regions, but the tiny particles called *Aitken nuclei* can

travel across continents and even circumnavigate the Earth. Although condensation nuclei are essential for widespread condensation of water vapour, only a small fraction of the nuclei present in the air take part in cloud droplet formation at any one time.

Other conditions must be fulfilled before precipitation occurs. First, moist air must be cooled to near its dew point. This can be brought about in several ways as follows.

(a) By an adiabatic expansion of rising air. A volume of air may be forced to rise by an impeding mountain range. The reduction in pressure causes a lowering of temperature without any transference of heat.
(b) By a meeting of two very different air masses. For example, when a warm, moist mass of air converges with a cold mass of air, the warm air is forced to rise and may cool to the dew point. Any mixing of the contrasting masses of air would also lower the overall temperature.
(c) By contact between a moist air mass and a cold object such as the ground.

Once cloud droplets are formed, their growth depends on hygroscopic and surface tension forces, the humidity of the air, rates of transfer of vapour to the water droplets and the latent heat of condensation released. A large population of droplets competes for the available water vapour and so their growth rate depends on their origins and on the cooling rate of air providing the supply of moisture (Fig. 1.6).

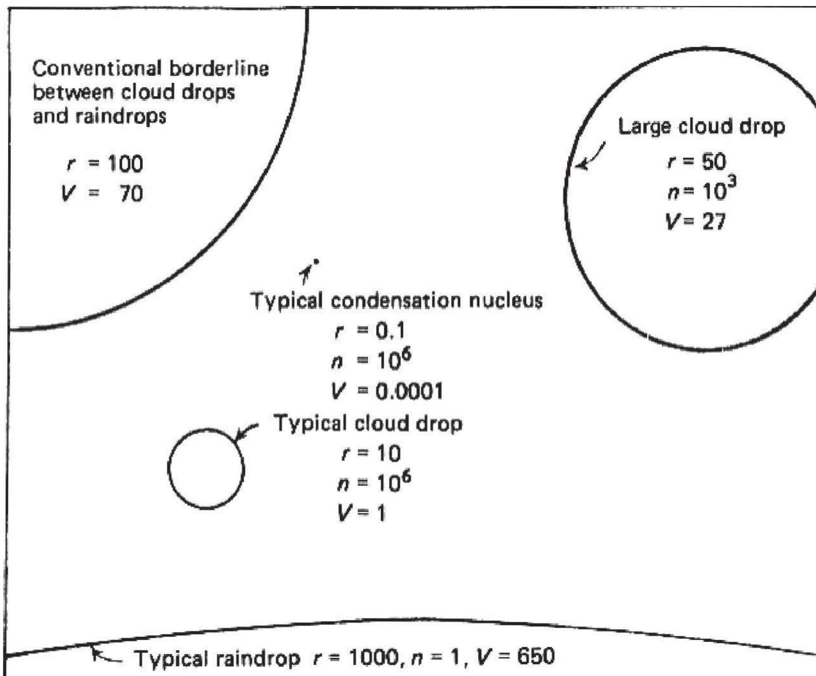


Fig. 1.6 Comparative sizes, concentrations and terminal falling velocities of some particles involved in condensation and precipitation processes, where r = radius (μm); n = number per dm^3 (10^3 cm^3); V = terminal velocity (cm s^{-1}). (Reproduced from B. J. Mason (1975) *Clouds, Rain and Rainmaking*, 2nd edn, by permission of Cambridge University Press.)

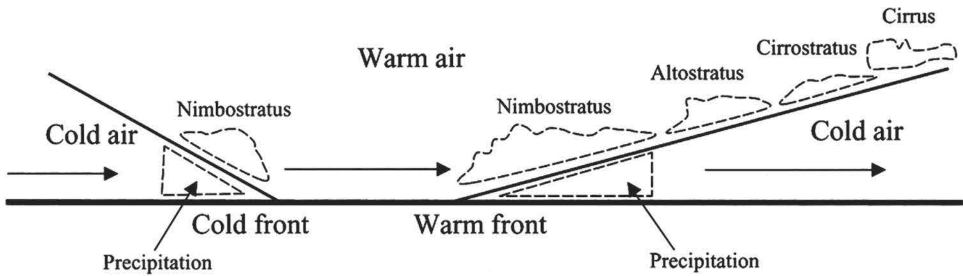


Fig. 1.7 Frontal weather conditions, showing cloud and precipitation around cold and warm fronts. (Reproduced with permission from I. Strangeways (2007) *Precipitation: Theory, Measurement and Distribution*, Cambridge University Press, Cambridge.)

The mechanism becomes complicated when the temperature reaches freezing point. Pure water can be supercooled to about −40° C (233 K) before freezing spontaneously. Cloud droplets are unlikely to freeze in normal air conditions until cooled below −10°C (263 K) and commonly exist down to −20°C (253 K). They freeze only in the presence of small particles called *ice nuclei*, retaining their spherical shape and becoming solid ice crystals. Water vapour may then be deposited directly on to the ice surfaces. The crystals grow into various shapes depending on temperature and the degree of supersaturation of the air with respect to the ice.

Condensed water vapour appears in the atmosphere as clouds in various characteristic forms; a standard classification of clouds is shown in Fig. 1.7. The high clouds are composed of ice crystals, the middle clouds of either water droplets or ice crystals, and the low clouds mainly of water droplets, many of them supercooled. Clouds with vigorous upwards vertical development, such as cumulonimbus, consist of cloud droplets in their lower layers and ice crystals at the top.

1.5.1 Theories of raindrop growth

Considerable research has been carried out by cloud physicists on the various stages involved in the transference of atmospheric water vapour into precipitable raindrops or snowflakes. A cloud droplet is not able to grow to raindrop size by the simple addition of water vapour condensing from the air. It is worth bearing in mind that one million droplets of radius 10 μm are equivalent to a single small raindrop of radius 1 mm. Fig. 1.6 shows the principal characteristics of nuclei, cloud droplets and raindrops.

Cloud droplets can grow naturally to about 100 μm in radius, and although tiny drops from 100 to 500 μm may, under very calm conditions, reach the ground, other factors are at work in forming raindrops large enough to fall to the ground in appreciable quantities. There are several theories of how cloud droplets grow to become raindrops, and investigations into the details of several proposed methods continue to claim the attention of research workers.

The Bergeron process, named after the famous Norwegian meteorologist, requires the coexistence in a cloud of supercooled droplets and ice particles and a temperature less than 0° C (273 K). The air is saturated with respect to water but supersaturated with respect to ice. Hence water vapour is deposited on the ice particles to form ice crystals.

The air then becomes unsaturated with respect to water so droplets evaporate. This process continues until either all the droplets have evaporated or the ice crystals have become large enough to drop out of the cloud to melt and fall as rain as they reach lower levels. Thus the crystals grow at the expense of the droplets. This mechanism operates best in clouds with temperatures in the range −10 to −30°C (263–243 K) with a small liquid water content.

Growth by collision: in clouds where the temperature is above 0°C (273 K), there are no ice particles present and cloud droplets collide with each other and grow by coalescence. The sizes of these droplets vary enormously and depend on the size of the initial condensation nuclei. Larger droplets fall with greater speeds through the smaller droplets with which they collide and coalesce. As larger droplets are more often formed from large sea-salt nuclei, growth by coalescence operates more frequently in maritime than in continental clouds. In addition, as a result of the dual requirements of a relatively high temperature and generous liquid water content, the growth of raindrops by coalescence operates largely in summer months in low-level clouds.

When cloud temperatures are below 0°C (273 K) and the cloud is composed of ice particles, their collision causes growth by *aggregation* to form snowflakes. The most favourable clouds are those in the 0 to −4°C (269 K) range and the size of snowflakes decreases with the cloud temperature and water content.

Growth by accretion occurs in clouds containing a mixture of droplets and ice particles. Snow grains, ice pellets or hail are formed as cloud droplets fuse on to ice particles. Accretion takes place most readily in the same type of cloud that favours the Bergeron process, except that a large content of liquid water is necessary for the water droplets to collide with the ice particles.

Even when raindrops and snowflakes have grown large enough for their gravity weight to overcome up-draughts of air and fall steadily towards the ground, their progress is impeded by changing air conditions below the clouds. The temperature may rise considerably near the Earth's surface and the air may become unsaturated. As a result snowflakes usually melt to raindrops and the raindrops may evaporate in the drier air. On a summer's day it is not uncommon to see cumulus clouds trailing streams of rain which disappear before they reach the ground. With dry air below a high cloud base of about 3 km, all precipitation will evaporate. Hence it is rare to see rainfall from altocumulus, altostratus and higher clouds (see Fig. 1.7). Snowflakes rarely reach the ground if the surface air temperature is above 4°C, but showers of fine snow can occur with the temperature as high as 7°C, if the air is very dry.

Further explanation of the processes involved in raindrop formation is given in Sumner (1988) and Strangeways (2007).

1.6 Weather patterns producing precipitation

The main concern of the meteorologist is an understanding of the general circulation of the atmosphere with the aim of forecasting the movements of pressure patterns and their associated winds and weather. It is sufficient for the hydrologist to be able to identify the situations that provide the precipitation, and for the practising civil engineer to keep a 'weather eye' open for adverse conditions that may affect his site work.

The average distribution and seasonal changes of areas of high atmospheric pressure (*anticyclones*) and of low-pressure areas (*depressions*) can be found in most

good atlases. Associated with the location of anticyclones is the development of homogeneous air masses. A *homogeneous air mass* is a large volume of air, generally covering an area greater than 1000 km in diameter, which shows little horizontal variation in temperature or humidity. It develops in the stagnant conditions of a high-pressure area and takes on the properties of its location (known as a *source region*). In general, homogeneous air masses are either cold and stable, taking on the characteristics of the polar regions from where they originate, or they are warm and unstable, revealing their tropical source of origin. Their humidity depends on whether they are centred over a large continent or over the ocean. The principal air masses are summarized in Table 1.6. Differences in atmospheric pressure cause air masses to move from high- to low-pressure regions and they become modified by the environments over which they pass. Although they remain homogeneous, they may travel so far and become so modified that they warrant reclassification. For example, when polar maritime air reaches the British Isles from a south-westerly direction, having circled well to the south over warm subtropical seas, its character will have changed dramatically.

Precipitation can come directly from a maritime air mass that cools when obliged to rise over mountains in its path. Such precipitation is known as *orographic rainfall* (or snowfall, if the temperature is sufficiently low), and is an important feature of the western mountains of the British Isles, which lie across the track of the prevailing winds bringing moisture from the Atlantic Ocean. Orographic rain falls similarly on most hills and mountains in the world, with similar locational characteristics, though it may occur only in particular seasons.

When air is cooled as a result of the converging of two contrasting air masses, it can produce more widespread rainfall independent of surface land features. The boundary between two air masses is called a *frontal zone*. It intersects the ground at the *front*, a band of about 200 km across. The character of the front depends on the difference between the air masses. A steep temperature gradient results in a strong or *active front* and much rain, but a small temperature difference produces only a *weak front* with less or even no rain. The juxtaposition of air masses across a frontal zone gives rise to two principal types of front according to the direction of movement.

Fig. 1.8 illustrates cloud patterns and the weather associated firstly with a *warm front*, in which warm air is replacing cold air, and secondly with a *cold front*, in which cold air is pushing under a warm air mass. In both cases, the warm air is made to rise and hence cool, and the condensation of water vapour forms characteristic clouds

Table 1.6 Classification of air masses

Air mass	*Source region*	*Properties of source*
Polar maritime (Pm)	Oceans; 50° latitude	Cool, rather moist, unstable
Polar continental (Pc)	Continents in vicinity of Arctic Circle; Antarctica	Cool, dry, stable
Arctic or Antarctic (A)	Arctic Basin and Central Antarctica in winter	Very cold, dry, stable
Tropical maritime (Tm)	Sub-tropical oceans	Warm and moist; unstable inversion common feature
Tropical continental (Tc)	Deserts in low latitude; primarily the Sahara and Australian deserts	Hot and dry

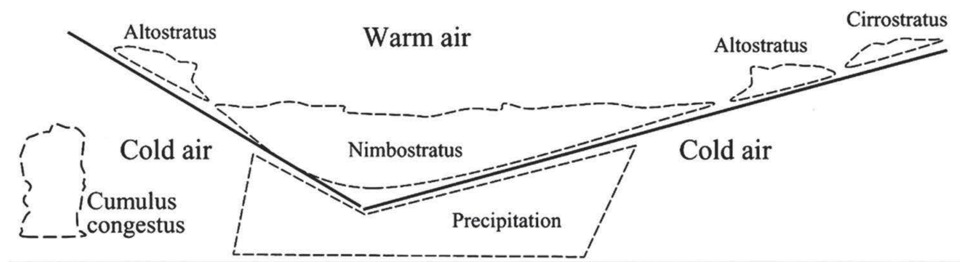


Fig. 1.8 Frontal weather conditions, showing cloud and precipitation development around an occluded front. (Reproduced from Strangeways, I. (2007) *Precipitation: Theory, Measurement and Distribution*, Cambridge University Press, Cambridge.)

and rainfall. The precipitation at a warm front is usually prolonged with gradually increasing intensity. At a cold front, however, it is heavy and short-lived. Naturally, these are average conditions; sometimes no rain is produced at all.

Over the world as a whole there are distinctive regions between areas of high pressure where differing air masses confront each other. These are principally in the mid-latitudes between 30° and 60° in both hemispheres, where the main boundary, the *polar front*, separates air masses having their origins in polar regions from the tropical air masses.

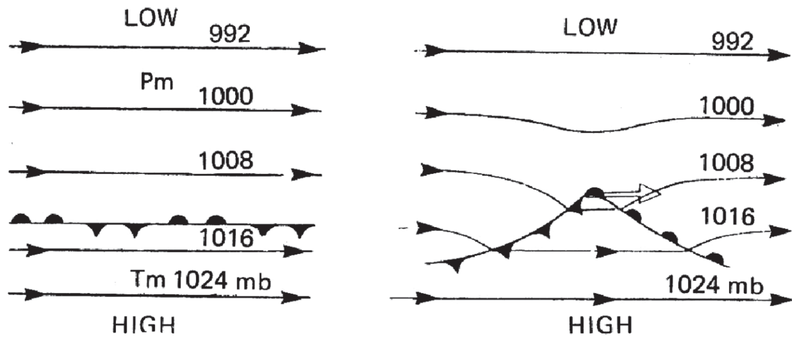
In addition, there is a varying boundary between air masses originating in the northern and southern hemispheres known as the *intertropical convergence zone* (ITCZ). The seasonal migration of the ITCZ plays a large part in the formation of the monsoon rains in south-east Asia and in the islands of Indonesia.

Four major weather patterns producing precipitation have been selected for more detailed explanation.

1.6.1 Mid-latitute cyclones or depressions

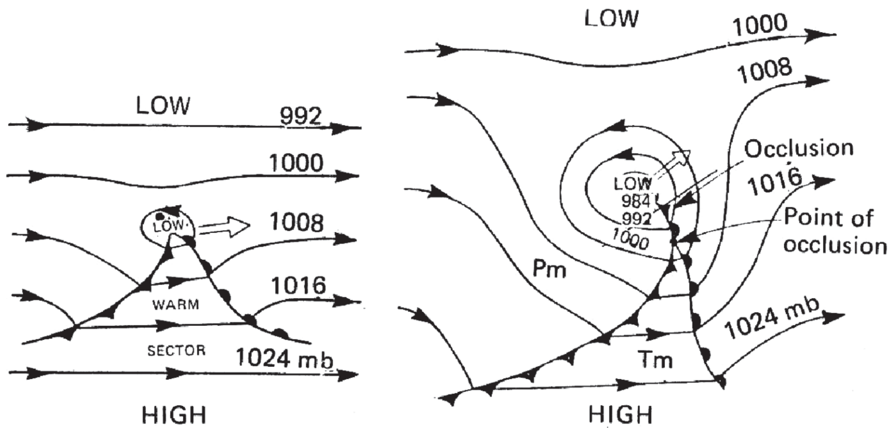
Depressions are the major weather pattern for producing precipitation in the temperate regions. More than 60 per cent of the annual rainfall in the British Isles comes from such disturbances and their associated features. They develop along the zone of the polar front between the polar and tropical air masses. Knowledge of the growth of depressions, the recognition of air masses and the definition of fronts all owe much to the work of the Norwegian meteorologists Wilhelm and Jacob Bjerknes in the 1920s.

The main features in the development and life of a mid-latitude cyclone are shown in Fig. 1.9. The first diagram illustrates in plan view the isobars of a steady-state condition at the polar front between contrasting air masses. The succeeding diagrams show the sequential stages in the average life of a depression. A slight perturbation caused by irregular surface conditions, or perhaps a disturbance in the lower stratosphere, results in a shallow wave developing in the frontal zone. The initial wave, moving along the line of the front at 15–20 m s^{-1} (30–40 knots), may travel up to 1000 km without further development. If the wavelength is more than 500 km, the wave usually increases in amplitude, warm air pushes into the cold air mass and active fronts are formed.



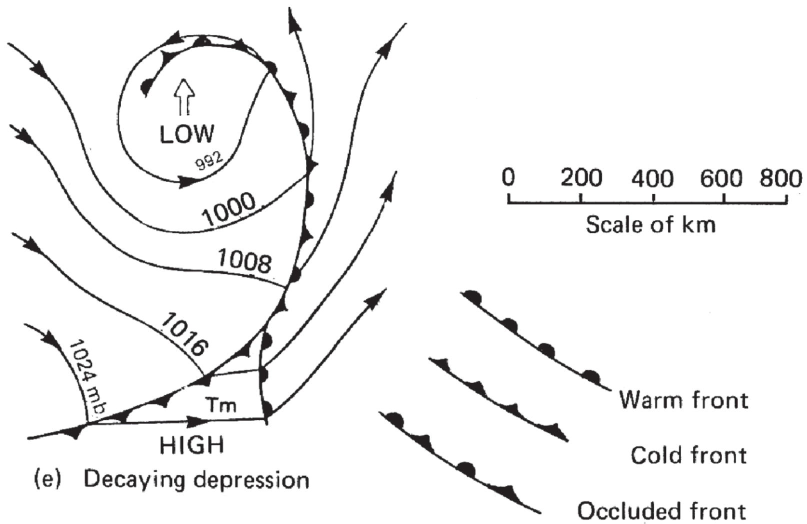
(a) Quasi-stationary part of the polar front

(b) Frontal wave



(c) Warm-sector depression

(d) Partly occluded depression



(e) Decaying depression

Fig. 1.9 Life cycle of a model occluding depression. (Adapted from Met Office (1962) *A Course in Elementary Meteorology*, Her Majesty's Stationary Office.)

As a result, the air pressure falls and a 'cell' of low pressure becomes trapped within the cold air mass. Gradually the cold front overtakes the warm front, the warm air is forced aloft, and the depression becomes *occluded*. The low-pressure centre then begins to fill and the depression dies as the pressure rises. On average, the sequence of growth from the first perturbation of the frontal zone to the occlusion takes 3–4 days. Precipitation usually occurs along the fronts and, in a very active depression, large amounts can be produced by the occlusion, especially if its speed of passage is retarded by increased friction at the Earth's surface. At all stages, orographic influences can increase the rainfall as the depression crosses land areas. A range of mountains can delay the passage of a front and cause longer periods of rainfall. In addition, if mountains delay the passage of a warm front, the occlusion of the depression may be speeded up.

1.6.2 Waves in the easterlies and tropical cyclones

Small disturbances are generated in the trade wind belts in latitudes 5–25° both north and south of the Equator. Irregular wind patterns showing as isobaric waves on a weather map develop in the tropical maritime air masses on the equatorial side of the subtropical high-pressure areas. They have been studied most in the Atlantic Ocean to the north of the South American continent. A typical easterly wave is shown in Fig. 1.10. A trough of low pressure is shown moving westwards on the southern flanks of the Azores anticyclone. The length of the wave extends over 15–20° longitude (1500 km) and, moving with an average speed of 6.7 m s^{-1} (13 knots), takes 3–4 days to pass. The weather sequence associated with the wave is indicated beneath

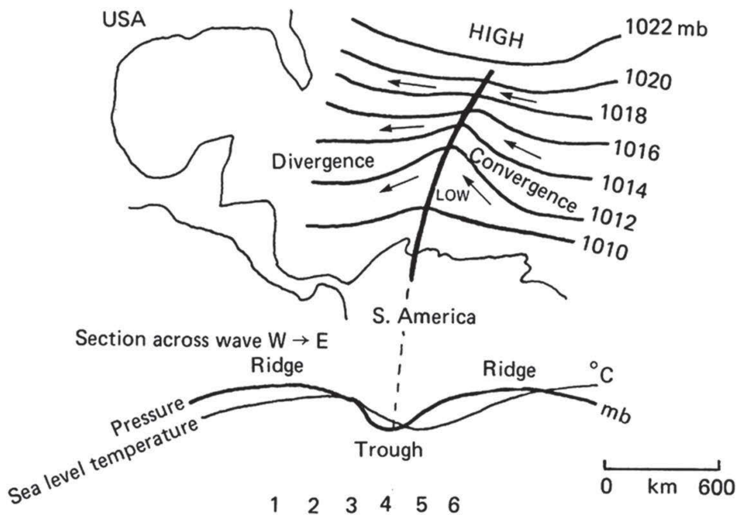


Fig. 1.10 A wave in the easterlies. Weather sequence: 1 – small Cu, no pp.; 2 – Cu, a few build-ups, haze, no pp.; 3 – larger Cu, Ci and Ac, better visibility, pr … pr … pr (showers); 4 – very large Cu, overcast Ci Ac, prpr or rr (continuous rain); 5 – Cu and Cb, Sc, As, Ac, Ci, pRpR (heavy showers), (thunderstorm); 6 – large Cu, occasional Cb, some Sc, Ac, Ci, pRpr – prpr; Cu = Cumulus; Cb = Cumulonimbus; Ci = Cirrus; Ac = Altocumulus; As = Altostratus; Sc = Stratucumulus.

the diagram. In the tropics, the cloud-forming activity from such disturbances is vigorous and subsequent rainfall can be very heavy: up to 300 mm may fall in 24 h.

As in mid-latitudes, the wave may simply pass by and gradually die away, but the low pressure may *deepen* with the formation of a closed circulation with encircling winds. The cyclonic circulation may simply continue as a shallow depression giving increased precipitation but nothing much else. However, rapidly deepening pressure below 1000 mb usually generates hurricane-force winds blowing round a small centre of 30–50 km radius, known as the *eye*. At its mature stage, a *hurricane* centre may have a pressure of less than 950 mb. Eventually the circulation spreads to a radius of about 300 km and the winds decline. Copious rainfall can occur with the passage of a hurricane; record amounts have been measured in the region of Southeast Asia, where the effects of the storms have been accentuated by orography. However, the rainfall is difficult to measure in such high winds. In fact, slower moving storms usually give the higher records. Hurricanes in the region of Central America often turn northwards over the United States and die out over land as they lose their moisture. On rare occasions disturbances moving along the eastern coastal areas of the United States are carried into westerly air-streams and become vigorous mid-latitude depressions.

Hurricanes tend to be seasonal events occurring in late summer when the sea temperatures in the areas where they form are at a maximum. They are called *typhoons* in the China Seas and *cyclones* in the Indian Ocean and off the coasts of Australasia. These tropical disturbances develop in well-defined areas and usually follow regular tracks; an important fact when assessing extreme rainfalls in tropical regions (McGregor and Nieuwolt, 1998).

1.6.3 Convectional precipitation

A great deal of the precipitation in the tropics is caused by local conditions that cannot be plotted on the world's weather maps. When a tropical maritime air mass moves over land at a higher temperature, the air is heated and forced to rise by convection. Very deep cumulus clouds form, becoming cumulonimbus extending up to the tropopause. Fig. 1.11 shows the stages in the life cycle of a typical cumulonimbus. Sometimes these occur in isolation, but more usually several such convective cells grow together and the sky is completely overcast.

The development of convective cells is a regular daily feature of the weather throughout the year in many parts of the tropics, although they do not always provide rain. Cumulus clouds may be produced but evaporate again when the air ceases to rise. With greater vertical air velocities, a large supply of moisture is carried upwards. As it cools to condensation temperatures, rainfall of great intensity occurs. In extreme conditions, hail is formed by the sequential movement of particles up and down in the cloud, freezing in the upper layers and increasing in size by gathering up further moisture. As the rain and hail fall, they cause vigorous down draughts, and when these exceed the vertical movements, the supply of moisture is reduced, condensation diminishes and precipitation gradually dies away. Thunder and lightning are common features of convectional storms with the interaction of opposing electrical charges in the clouds. The atmospheric pressure typically is irregular during the course of a storm.

Convectional activity is not confined to the tropics; it is a common local rain-forming phenomenon in higher latitudes, particularly in the summer. Recent studies have shown

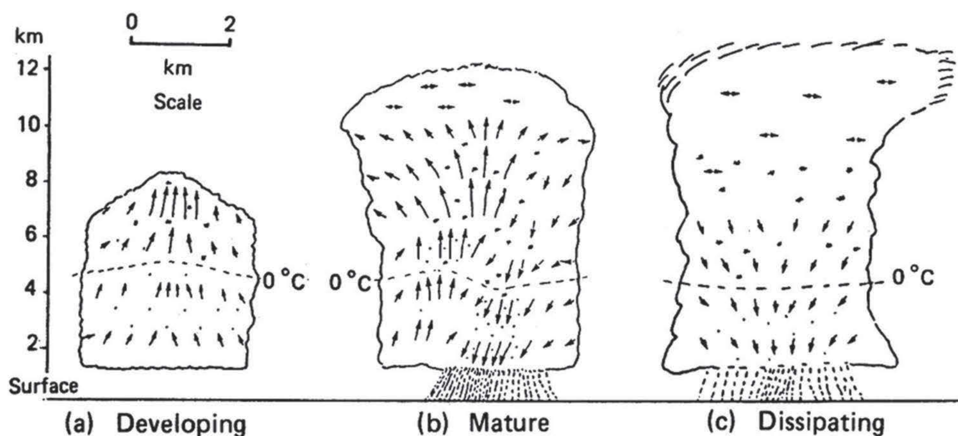


Fig. 1.11 Convective cells – stages in the life cycle. Time scales: (a) approximately 20 min; (b) approximately 20 min, heavy rain and hail, thunder may develop; (c) 30 min to 2 h, rainfall intensity decreasing. Total life cycle 1–2 h.↔, ice; *, snow; · · · ·, rain and hail; ↑↓, winds.

that convection takes place along frontal zones thus adding to rainfall intensities. Wherever strong convectional forces act on warm moist air, rain is likely to form and it is usually of high intensity over a limited area.

1.6.4 Monsoons

Monsoons are weather patterns of a seasonal nature caused by widespread changes in atmospheric pressure. The most familiar example is the monsoon of Southeast Asia where the dry, cool or cold winter winds blowing outwards from the Eurasian anticyclone are replaced in summer by warm or hot winds carrying moist air from the surrounding oceans being drawn into a low-pressure area over northern India. The seasonal movements of the ITCZ play a large part in the development and characteristics of the weather conditions in the monsoon areas. The circulation of the whole atmosphere has a direct bearing on the migration of the ITCZ, but in general the regularity of the onset of the rainy seasons is a marked feature of the monsoon. Precipitation, governed by the changing seasonal winds, can be caused by confrontation of differing air masses, low-pressure disturbances, convection and orographic effects. A map of the monsoon areas is shown in Fig. 1.12. Actual quantities of rain vary, but as in most tropical and semi-tropical countries, intensities are high (McGregor and Nieuwolt, 1998).

Further explanation of the processes producing the weather systems described is given in Holton (2004).

1.7 Climate

Following the appreciation of the meteorological mechanisms that affect evaporation and produce precipitation, it is pertinent to consider these hydrological processes

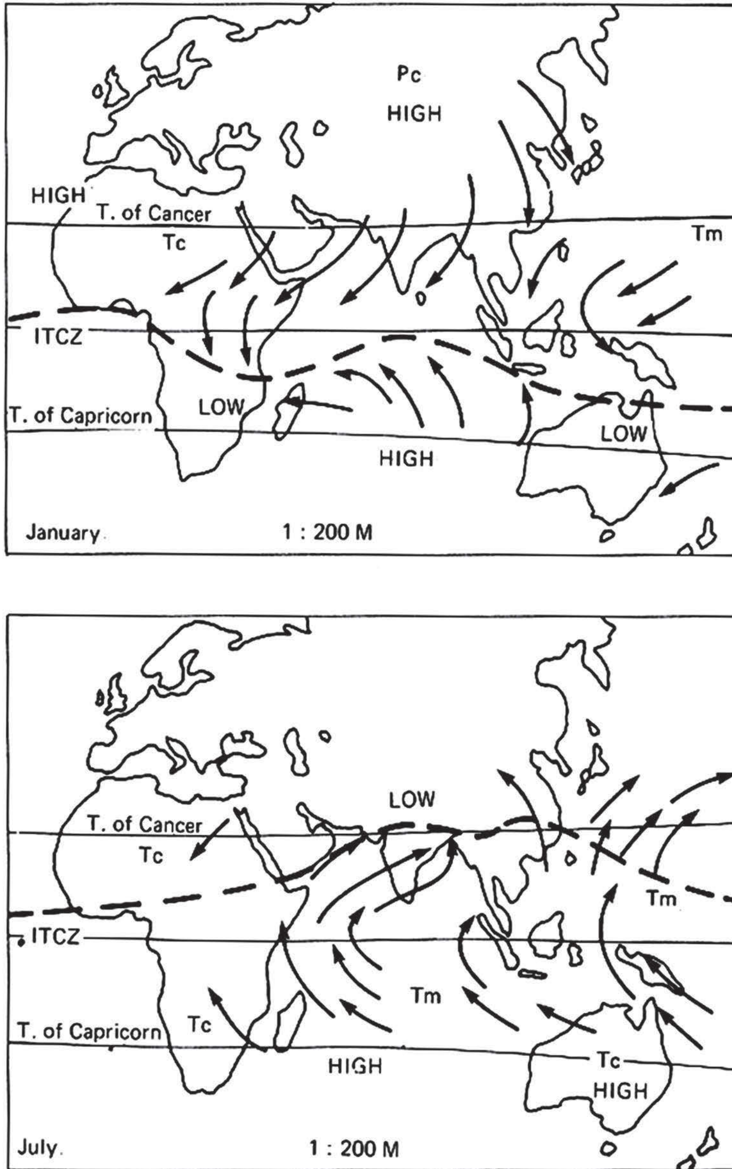


Fig. 1.12 Monsoon lands – pressure systems and winds.

on a longer time scale. Evaporation was presented as an instantaneous process. The precipitation-forming mechanisms extended into weather patterns that may last up to about a week. The study of climate is based on *average weather conditions*, specified usually by measures of temperature and precipitation over one or more months though other phenomena may also be aggregated. Statistics gathered for each month over a period of years and averaged give a representation of the climate of the location.

Table 1.7 Köppen climate classification

		Estimated percentages	
		Land surfaces	*Total surface*
A	Tropical rain climates – forests	20	36
B	Arid climates	26	11
C	Warm temperate rain climates – trees	15	27
D	Boreal forest and snow climates	21	7
E	Treeless cold snow climates	17	19

The most renowned classification of climates is that of Köppen who categorized climates according to their effect on vegetation.

The major groupings are given in Table 1.7. Subdivisions of these main groups are defined by thresholds of temperature and rainfall values; the details are given in most books of climatology. Their geographical distribution is shown in Fig. 1.13, though Peel *et al.* (2007) provide a revised (colour) version of this map.

These broad definitions of climatic regions are built up from the instrumental records of observing stations which are thus providing sample statistics representing conditions over varying areas. Such meteorological records have only been made with any reliability since the advancing development of instruments in the seventeenth century (Manley, 1970), and world coverage was limited until the late nineteenth century.

Before instrumental records, knowledge of the climate of different regions has been built up by the study of what is now called proxy data. For example, in the UK the proportion of certain tree pollens found in layers of lake sediments or upland peats give indications of the existence of tree cover in earlier times. Similarly, varying layers of clays and silts in surface deposits, as in Sweden, help to differentiate between warm and cold periods. In the western United States, the study of growth rings in the trunks of very old trees, allow climatologists to extend climatic information to periods before instrumental observations. On the global scene, the analyses of deep-sea sediments and ice cores are of increasing importance in the assemblage of climate knowledge.

In addition, archaeological and historical records of transient events, such as the extent of sea ice round the Poles, the fluctuation of mountain glaciers and even the variation of man's activities in the extent of vine growing and the abundance of the wheat harvests, all contribute clues to the climate of former times.

The assimilation and interpretation of such variable information gathered worldwide has occupied climatologists for many years and a broadly agreed sequence of climatic events has been established, aided by the findings of the geologists. However, the worldwide coverage of climatic information before this century was far from representative of all land regions and even less was known of the much larger oceanic areas. The recent concern over man-made changes in the composition of the atmosphere and the increasing ability to model changes in climate has led scientists to study the dynamic components contributing to climate as distinct from current weather, and the resultant impacts on hydrology (Bates *et al.*, 2008). Detailed discussion of climate change impacts on hydrology are presented within Chapter 19.

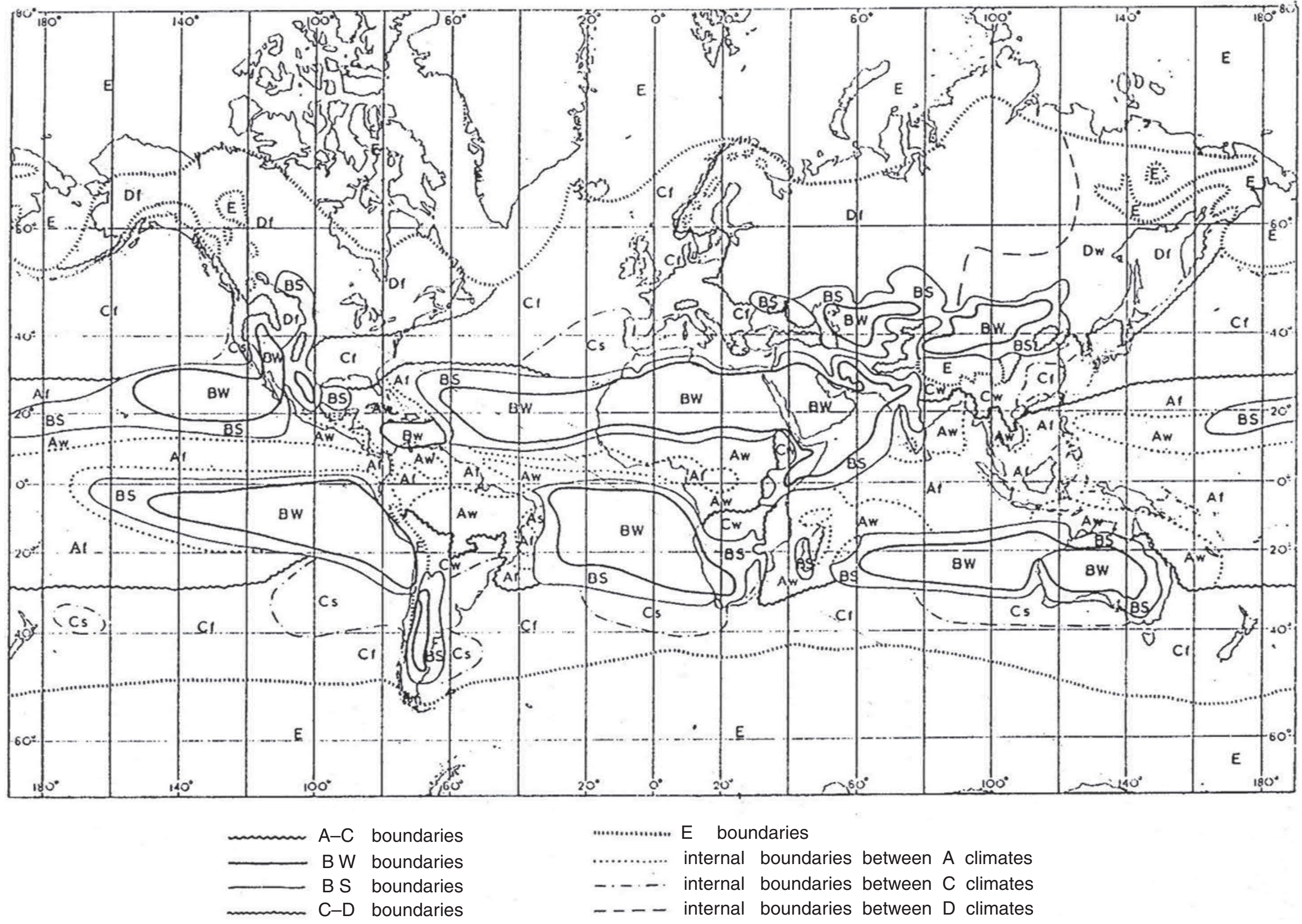


Fig. 1.13 Köppen's world classification of climates.

Note

1 Runoff is the river discharge ($m^3\,s^{-1}$) per unit catchment area, hence has the units of $m\,s^{-1}$ or $mm\,h^{-1}$.

References

Bates, B. C., Kundzewicz, Z. W., Wu, S. and Palutikof, J. P. (eds) (2008) *Climate Change and Water*. Technical Paper of the Intergovernmental Panel on Climate Change, IPCC Secretariat, Geneva, 210 pp.

Beven, K. J. (2004) Robert Horton's perceptual model of infiltration. *Hydrological Processes* 18, 3447–3460.

Beven, K. J. (2006) *Benchmark Papers in Storm Runoff Generation*. IAHS Press, Wallingford.

Biswas, A. K. (1970) *History of Hydrology*. North-Holland Pub. Co., Amsterdam, 336 pp.

Brustseart, W. (2005) *Hydrology – An Introduction*. Cambridge University Press, Cambridge.

Bunting, B. T. (1961) The role of seepage moisture in soil formation, slope development and stream initiation. *American Journal of Science* 259, 503–518.

Dooge, J. C. I. (1959) *Un bilan hydrologique au XVIIe siecle*. Houille Blanche, 14e annee No. 6, 799–807.

Dunne, T. (1978) Field studies of hillslope flow processes. In: Kirkby, M. J. (ed.) *Hillslope Hydrology*, John Wiley & Sons, pp. 227–294.

Gleick, P. H. (1996) Water resources. In: Schneider, S. H. (ed.) *Encyclopedia of Climate and Weather*, Vol. 2. Oxford University Press, New York, pp. 817–823.

Holton, J. R. (2004) *An Introduction to Dynamical Meteorology*, 4th edn. Academic Press, San Diego, 535 pp.

Manley, G. (1970) The climate of the British Isles. In: Wallen, C. C. (ed.) *Climates of Northern and Western Europe*, Elsevier, Amsterdam, 81–133. (*World Survey of Climatology*, Vol. 5)

Mason, B. J. (1975) *Clouds, Rain and Rainmaking*, 2nd edn. Cambridge University Press, Cambridge, 189 pp.

McGregor, G. R. and Nieuwolt, S. (1998) *Tropical Climatology: An Introduction to the Climates of the Low Latitudes*. Wiley, Chichester, 339 pp.

Met Office (1978) *A Course in Elementary Meteorology*, 2nd edn. HMSO, London, 208 pp.

Peel, M. C., Finlayson, B. L. and McMahon, T. A. (2007). Updated world map of the Köppen–Geiger climate classification. *Hydrology and Earth System Science* 1, 1633–1644.

Sklash, M. G. and Farvolden, R. N. (1979) The role of groundwater in storm runoff. *Journal of Hydrology* 43, 45–65.

Sklash, M. G., Beven, K. J., Gilman, K. and Darling, W. G. (1996) Isotope studies of pipeflow in Plynlimon, Wales, UK. *Hydrological Processes* 10, 921–944.

Strangeways, I. (2007) *Precipitation: Theory, Measurement and Distribution*. Cambridge University Press, Cambridge.

Sumner, G. (1988) *Precipitation: Process and Analysis*. John Wiley & Sons, Chichester, 455 pp.

Weiler, M. and McDonnell, J. (2004) Virtual experiments: a new approach for improving process conceptualization in hillslope hydrology. *Journal of Hydrology* 285, 3–18.

Weyman, D. R. (1970) Throughflow on hillslopes and its relation to the stream hydrograph. *International Association of Scientific Hydrology Bulletin* 15, 25–33.

Chapter 2

Hydrometric networks

The concept of the hydrological cycle forms the basis for the hydrologist's understanding of the sources of water at or under the Earth's surface and its consequent movement by various pathways back to the principal storage in the oceans. Two of the greatest problems for the hydrologist are quantifying the amount of water in the different phases in the cycle and evaluating the rate of transfer of water from one phase to another within the cycle. Thus measurement of the components of the cycle is a major objective of the engineering hydrologist and scientific hydrologist.

Nationwide schemes to measure hydrological variables are now considered essential for the development and management of the water resources of a country. As a result, responsibility for measurement stations is focused on central or regional government agencies and detailed considerations are afforded to the planning of hydrological measurements. Cost–benefit assessments are also being made on the effectiveness of data gathering, and hence scientific planning is being recommended to ensure optimum networks to provide the required information.

2.1 Gauging networks

One of the main activities stimulated by the International Hydrological Decade (IHD, 1965–1974) was the consideration of hydrological network design, a subject that, it was felt, had been previously neglected (Rodda, 1969). It was recognized that most networks, even in developed countries, were inadequate to provide the data required for the increasing need of hydrologists charged with the task of evaluating water resources for expanding populations. While the situation has improved considerably within developed countries, the poor coverage of hydrometric networks within many tropical countries remains a concern.

Before approaching the problem, it is pertinent to define a network. Langbein (1965) gave a broad definition: 'A network is an organized system for the collection of information of a specific kind: that is, each station, point or region of observation must fill one or more definite niches in either space or time'.

The design of the optimum hydrometric network must be based on quantified objectives wherever possible, with costs and benefits included in the design procedure. One approach is the evaluation of the worth of the data collected, which sometimes means realizing the benefits lost through lack of data. Closely connected with network design and data collection is an appreciation of the errors within hydrometric data (Herschy, 1999).

2.2 Design considerations

There are several well-defined stages in the design of a network of gauging stations for the measurement of hydrological variables. The first comprises initial background research on the location and known characteristics of the area to be studied. The size of the area and whether it is a political entity or a natural drainage basin are of prime importance. When assessing the design problem, it is advisable to think in terms of natural catchment areas even if the total area is defined by political boundaries. The physical features of the area should be studied. These include the drainage pattern, the surface relief (altitudinal differences), the geological structure and the vegetation. The general features of the climate should be noted; seasonal differences in temperature and precipitation can be identified from good atlases or standard climatological texts. The characteristics of the precipitation also affect network design and the principal meteorological causes of the rainfall or snowfall should be investigated.

The second stage in network design involves the practical planning. Existing measuring stations should be identified, visited for site inspection and to determine observational practice, and all available data assembled. The station sites should be plotted on a topographical map of the area or, if the area is too large, one overall locational map should be made and separate topographical maps compiled for individual catchment areas. The distribution of the measuring stations should be studied with regard to physical features and data requirements, and new sites chosen to fill in any gaps or provide more detailed information for special purposes. The number of new gauging sites required depends on the density of stations considered to be an optimum for the area. (Indications of desirable station densities are given in the following sections.) Any new sites in the network are chosen on the map, but then they must be identified on the ground. Visits to proposed locations are essential for detailed planning and selected sites may have to be adjusted to accord with ground conditions.

The third stage involves the detailed planning and design of required installations on the new sites. These vary in complexity according to the hydrological variable to be measured, ranging from the simple siting of a single storage rain gauge to the detailed designing of a compound weir for stream measurements or the drilling of boreholes for monitoring groundwater levels. The costing of the hydrometric scheme is usually done at this stage and when this is approved and the finance is available, steps can be taken to execute the designed scheme.

A procedure that may be carried out at any stage is the testing of the validity of the data produced by the network with or without any new stations, provided that there are enough measurements available from the existing measurement stations to allow significant statistical analyses. These may take various forms depending on the variability of the measurements being tested. The worth of data produced is now an important factor in network design, but such cost–benefit evaluation is complicated by the many uses made of the data and by the unknown applications that may arise in the future.

The ideal hydrometric scheme includes plans for the measurement of all the many different hydrological variables, including water quality. Designed networks of water quality monitoring stations are now being established in conjunction with arrangements for measurements of quantity. In the following sections, further particulars of

network design for the more usual variables, precipitation, evaporation, overland flow, subsurface flow, river flow and groundwater will be given.

2.3 Precipitation networks

The design of a network of precipitation gauging stations is of major importance to the hydrologist since it is intended to provide a measure of the water input to the river catchment system. The rainfall input is irregularly distributed both over the catchment area and in time. Another consideration in precipitation network design must be the rainfall type as demonstrated in the areal rainfall errors obtained over a catchment of 500 km^2 having ten gauges (Table 2.1). This also shows that a higher density of gauges is necessary to give acceptable areal values on a daily basis.

As a general guide to the density of precipitation stations required, Table 2.2 gives the absolute minimum density for different parts of the world. The more variable the areal distribution of precipitation, as in mountainous areas, the more gauges are needed to give an adequate sample. In regions of low rainfall totals, the occurrence is variable but the infrequent rainfall events tend to be of higher intensity and thus network designers should ensure adequate sampling over areas that would be prone to serious flooding (see Chapter 16).

In the UK, recommended minimum numbers of rain gauges for reservoired moorland areas were laid down by water engineers many years ago (Institute of Water Engineers, 1937; Table 2.3). For real-time operation of an upland impounding reservoir at least one recording gauge would now be recommended to record heavy falls over short periods.

From measurement theory for any random variable, the recommended number of rain gauges should be based on the standard error in rainfall for the particular location

Table 2.1 Areal rainfall errors (%). Reproduced from J.C. Rodda (1969) *Hydrological Network Design – Needs, Problems and Approaches*. WMO/IHD Report No. 12.

Type of rainfall	*Day*	*10 days*	*Month*	*Season*
Frontal	19	8	4	2
Convective	46	17	10	4

Table 2.2 Minimum density of precipitation stations. (Reproduced from World Meterological Organization (1965) *Guide to Hydrometeorological Practices*)

Region	*Minimum density range (km^2/gauge)*
Temperate, Mediterranean and tropical zones	
Flat areas	600–900
Mountainous areas	100–250
Small mountainous islands (<20 000 km^2)	25
Arid and polar zones	1500–10 000

Table 2.3 Rain gauge networks for the UK: minimum numbers of rain gauges required in reservoired moorland areas. (Reproduced from Institute of Water Engineers (1937) *Transactions of the Institute of Water Engineers* XLII, 231–259)

Area (km^2)	*Rain gauges*		
	Daily	*Monthly*	*Total*
2	1	2	3
4	2	4	6
20	3	7	10
41	4	11	15
81	5	15	20
122	6	19	25
162	8	22	30

and sampling period. The standard error of the mean, σ_{err}, may be used to estimate the closeness of the sample mean to the true mean,

$$\sigma_{err} = \frac{\sigma_s}{\sqrt{n}}$$

where σ_s is the standard deviation, and n is the number of independent observations. Thus, if the rainfall is expected to have a standard deviation of 50 mm, and we wish to approximate the population mean to a standard error of 10 mm, then the number of rain gauges required is 25. This is, however, an underestimate of the minimum gauge density if the spatial pattern of rainfall is sought (rather than just a catchment average). For this, an estimate of the geostatistical structure of the rainfall field could be used to derive the required rain gauge density (Moore *et al.*, 2000; Nour *et al.*, 2006; Villarini *et al.*, 2008; see also Section 9.8.1).

Ground-based radars are used within many nations to measure rainfall patterns. Each rainfall radar measures an area of approximately 15 000 km^2. The British Isles are currently covered by 16 rainfall radars, with a further four stations proposed (Holehead, Munduff Hill, Old Buckingham, High Moorsley; Section 3.6). The real-time combination of radar and rain gauge data can produce reasonable detail over very large areas (Alpuim and Barbosa, 1999) and is discussed further in Section 3.7.

2.4 Evaporation networks

The assessment of evaporation loss over a catchment by means of local measurements is the next to be considered. The various recommended methods of measurement are outlined in Chapter 4. Since evaporation and transpiration over an area are relatively conservative quantities in the hydrological cycle, fewer gauging stations are required to give areal evaporation estimates than for areal rainfalls. Evaporation and transpiration are dependent on altitude, and thus a network of measuring stations should sample different altitudinal zones within a catchment area. To give some idea of numbers, for experimental catchments covering 18 km^2 in Wales with a range in altitude of 460 m,

where 25 rain gauges are needed for evaluations of areal rainfall with a 2 per cent error, only three or four evaporation stations would be necessary for areal evaporation estimates (McCulloch, 1965). A reliable single station would provide adequate information over a flat plain or plateau. Variations could still occur, however, from differing types of vegetation.

Over 400 towers supporting sensors for measuring evaporation (Chapter 4) have been installed across the globe, with data shared across the FLUXNET network (Baldocchi *et al.*, 2001; Section 6.5). Further, the representativeness of these stations at describing evaporation across continental regions has been assessed recently (Yang *et al.*, 2008). Additionally, the data from this network have been combined with satellite data (e.g. MODIS temperature and humidity data) to derive regional estimates of evaporation (Cleugh *et al.*, 2007).

2.5 Overland flow networks

Networks of overland flow plots are important in the derivation of regional and national estimates of erosion rate. Estimates of the minimum number of overland flow troughs (Section 6.5) needed to characterise the variability in overland flow across landscapes can be obtained from the erosion literature (e.g. Nearing *et al.*, 1999).

2.6 Subsurface water networks

The main purpose of subsurface water investigation for water companies to identify the extent of productive *aquifers* (i.e. groundwater bodies with high porosity and high saturated hydraulic conductivity), to determine their hydraulic properties, and to make arrangements for monitoring the water levels within the aquifers. The sites of existing wells should be noted, as these may have water level records which give the long-term fluctuations of the water table. Furthermore, siting of boreholes for observations must take into account differences in hydraulic properties within an aquifer in addition to variations between aquifers. For example, in the UK, fairly homogeneous Triassic sandstone may require a basic network of one borehole per 260 km^2. However, chalk aquifers can be very variable in saturated hydraulic conductivity and consequently a denser network of boreholes, say 1 in 5 km^2, may be needed to record water level fluctuations (Ineson, 1965). A more comprehensive overview of the design criteria for subsurface monitoring is given in Sara (2006), including the use of geostatistical methods (Carrera *et al.*, 1984).

The monitored borehole network within the UK is detailed within Marsh and Hannaford (2008), and the importance of this network has been enhanced by the development of groundwater resources and the recharge of aquifers depleted by over-pumping.

2.7 River gauging networks

The establishment of river gauging stations is often the most costly item in a hydrometric scheme and, as such, river gauging is usually the responsibility of a national or regional authority, e.g. the Environment Agency of England and Wales. The density

of gauging stations depends on the nature of the terrain and for water resources on the population creating a water demand. In England and Wales, it was proposed that there should be 400 primary gauging stations, equivalent to a density of 1 in 375 km^2 (Boulton, 1965). When the Water Resources Act (1963) came into force, the number of gauging stations producing records for publication was approaching this figure, and coordination of further planning of the nationwide network was undertaken by the central authority, the Water Resources Board. The then individual River Authorities were advised on the status of the gauging stations required:

- primary or principal stations defined as permanent stations to measure all ranges of discharges and observations and records to be accurate and complete;
- secondary or subsidiary stations to operate for as long as necessary to obtain a satisfactory correlation with the record of a primary station; their function is to provide hydrological knowledge of streams likely to be used for water supply abstractions; the range of a secondary station should be as comprehensive as possible and the observations and records should be of primary station standard;
- special stations are those serving particular needs, such as reservoir levels and dry weather flow stations for controlling abstractions; these may be permanent or temporary stations according to requirements and they can be related to primary and secondary stations.

Currently within the UK, a combination of the Environment Agency, the Scottish Environmental Protection Agency, the Rivers Agency of Northern Ireland and many water companies maintain the network of (primary, secondary and special) river gauging stations and associated data. The UK Hydrometric Register (Marsh and Hannaford, 2008) currently reports data for around 1500 river gauging stations in the UK; this is equivalent to a density of 1 in 163 km^2.

For water resources evaluation, 20 years of records from a secondary station would suffice to give an acceptable correlation coefficient between the monthly discharges of the secondary station and a primary station. Then the secondary station could be discontinued. Extension of discharge information for a short-term secondary station can also be made by relating the *flow duration curves* (Section 11.4).

The ultimate design and establishment of a river gauging network depends on the data requirements, the hydrological characteristics of the area and the achievement of an acceptable cost–benefit relationship for the scheme.

There are increasing efforts to make data from river gauging stations available internationally. The Global Runoff Data Centre (GRDC),[1] maintained by the German Federal Institute of Hydrology (BfG), currently holds daily or monthly river-flow data for 7332 river gauging stations in 156 countries across the globe.

The following chapters of Part I dealing with hydrological measurements describe the methods of measurement of the different hydrological variables and the instruments in most common use in the UK.

Note

1 See http://www.bafg.de/GRDC/EN/Home/homepage__node.html

References

Alpuim, T. and Barbosa, S. (1999) The Kalman filter in the estimation of area precipitation. *Environmetrics*, 10, 377–394.

Baldocchi, D., Falge, E., Gu, L., Olson, R., Hollinger, D., Running, S., Anthoni, P., Bernhofer, Ch., Davis, K., Fuentes, J., Goldstein, A., Katul, G., Law, B., Lee, X., Malhi, Y., Meyers, T., Munger, J. W., Oechel, W., Pilegaard, K., Schmid, H. P., Valentini, R., Verma, S., Vesala, T., Wilson, K. and Wofsy, S. (2001) FLUXNET: a new tool to study the temporal and spatial variability of ecosystem-scale carbon dioxide, water vapour and energy flux densities. *Bulletin of the American Meteorological Society* 82, 2415–2435.

Boulton, A. G. (1965) *Surface Water, Basic Principles Related to Network Design*. IASH Pub. No. 67, 234–244.

Carrera, J., Usunoff, E. and Szidarovszky, F. (1984) A method for optimal observation network design for groundwater management. *Journal of Hydrology* 73, 147–163.

Cleugh, H. A., Leuning, R., Mu, Q. and Running, S. W. (2007) Regional evaporation estimates from flux tower and MODIS satellite data. *Remote Sensing of Environment* 106, 285–304 (doi: 10.1016/j.rse.2006.07.007).

Herschy, R. W. (1999) *Hydrometry: Principles and Practices,* 2nd edn. John Wiley, New York, 376 pp.

Ineson, J. (1965) *Ground Water Principles of Network Design*. IASH Pub. No. 68, 476–483.

Institute of Water Engineers (1937) Report of Joint Committee to Consider Methods of Determining General Rainfall Over Any Area, Transactions of the Institute of Water Engineers, XLII, 231–259.

Langbein, W. B. (1965) *National Networks of Hydrological Data*. IASH Pub. No. 67, 5–11.

Marsh, T. and Hannaford, J. (eds) (2008) UK Hydrographic Register. Hydrological data UK series. Centre for Ecology & Hydrology, Wallingford, 200pp.

McCulloch, J. S. G. (1965) *Hydrological Networks for Measurement of Evaporation and Soil Moisture*. IASH Pub. No. 68, 579–584.

Moore, R. J., Jones, D. A., Cox, D. R. and Isham, V. S. (2000) Design of the HYREX raingauge network. *Hydrology and Earth System Sciences* 4, 523–530.

Nearing, M. A., Govers, G. and Norton, L. D. (1999) Variability in soil erosion data from replicated plots. *Soil Science Society of America Journal* 63, 1829–1835.

Nour, M. H., Smit, D. W. and Gamal El-Din, M. (2006) Geostatistical mapping of precipitation: implications for rain gauge network design. *Water Science and Technology* 53, 101–110.

Rodda, J. C. (1969) *Hydrological Network Design – Needs, Problems and Approaches*. WMO/IHD Report No. 12, 57 pp.

Sara, M. N. (2006) Groundwater monitoring system design. In: Nielsen, D. M. (ed.) *Practical Handbook of Environmental Site Characterisation and Ground-Water Monitoring*, 2nd edn. Taylor & Francis, Boca Raton, 517–572.

Villarini, G., Mandapaka, P. V., Krajewski, W. F. and Moore, R. J. (2008) Rainfall and sampling uncertainties: a rain gauge perspective. *Journal of Geophysical Research - Atmospheres* 113, D11102. 12, pp. 10.1029/2007JD009214.

World Meteorological Organization (WMO; 1965) *Guide to Hydrometeorological Practices*. WMO, Geneva.

Yang, F., Zhu, A., Ichii, K., White, M. A., Hashimoto, H. and Nemani, R. R. (2008) Assessing the representativeness of the AmeriFlux network using MODIS and GOES data. *Journal of Geophysical Research* 113(g4), G04036.

Chapter 3

Precipitation

Of all the components of the hydrological cycle, the elements of precipitation, particularly rain and snow, are the most commonly measured. Sevruk and Klemm (1989) have estimated that there are 150 000 storage rain gauges in use worldwide. It would appear to be a straightforward procedure to catch rain as it falls and the depth of snow lying can be determined easily by readings on a graduated rod. People have been making these simple measurements for more than 2000 years; indeed, the first recorded mention of rainfall measurement came from India as early as 400 BC. The first rain gauges were used in Korea in the 1400s AD (as a means to plan farming and set taxes), and 200 years later, in ca. 1680 in England, Sir Christopher Wren and Robert Hooke described designs for the self-recording rain gauge.

Climatologists and water engineers appreciate that making an acceptable precipitation measurement is not as easy as it may first appear. It is not physically possible to catch all the rainfall or snowfall over a catchment; the precipitation over the area can only be *sampled* by rain gauges. The measurements are made at several selected points representative of the area and values of the total volume (Ml) or equivalent areal depth (mm) over the catchment are calculated later. Such are the problems in obtaining representative samples of the precipitation reaching the ground that, over the years, a comprehensive set of rules has evolved. The principal aim of these rules is to ensure that all measurements are comparable and consistent. All observers are recommended to use standard instruments installed uniformly in representative locations and to adopt regular observational procedures (as set within the particular country).

Many investigations carried out in England into the problems of rainfall measurement owe their origin to the enthusiasm of one man, G. J. Symons. Symons, a civil servant in the Meteorological Department of the Board of Trade in the 1850s, instigated and encouraged formal scientific experiments by such volunteers as retired army officers or clergymen whose spare time interests included observations of the weather and measurements of meteorological variables (Mill, 1901). The results of this work were incorporated by Symons into his *Rules for Rainfall Observers*. Symons' rules continue to form the basis of the practice of precipitation measurement in the UK today (Met Office, 2006).

The Met Office, which in 1919 inherited the advisory functions of Symons and his successors in the British Rainfall Organization, has approved instruments of several designs having the salient features recommended as a result of the early experiments. These include various types of storage rain gauge and automatic rain gauges. For the

assessment of water resources, monthly totals may suffice; for evaluating flood peaks in urban areas (Chapters 16 and 18), rainfall intensities over an hour or even minutes could be required, so automatic rain gauges are used.

3.1 Non-recording (storage) rain gauges

Rain gauges vary in capacity depending on whether they are to be read daily or monthly. The period most generally sampled is the day, and most precipitation measurements are the accumulated depths of water caught in simple storage gauges over 24 h.

For many years, the UK's recognized standard daily rain gauge has been the Met Office Mark II instrument (Fig. 3.1a; Met Office, 1980; British Standard 7843, 1996). The gauge has a sampling orifice of diameter 127 mm. The 12.7 mm rim is made of brass, the traditional material for precision instruments, and the sharply tooled knife edge defines a permanent accurate orifice. The Snowdon funnel forming the top part of the gauge has a special design. A straight-sided drop of 102 mm above the funnel prevents losses from out-splash in heavy rain. Sleet and light snowfall also collect readily in the deep funnel and, except in very low temperatures, the melted water runs down to join the rain in the collector. The Snowdon funnel, the main outer casing of the gauge and an inner can are all made of copper, a material that has a smooth surface, wets easily and whose surface, once oxidized, does not change. The inside of the collecting orifice funnel should never be painted, since the paint soon cracks, water adheres to the resulting rough surface and there are subsequent losses by evaporation. The main collector of the rain water is a glass bottle with a narrow neck to limit evaporation losses. The gauge is set into the ground with its rim level at 300 mm above the ground surface, which should ideally be covered with short grass, chippings or gravel to prevent in-splash in heavy rain.

During very wet weather, the rain collected in the bottle may overflow into the inner can. Bottle and can together hold the equivalent of 150 mm rainfall depth.

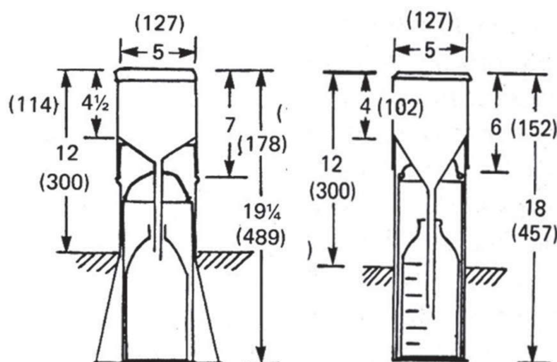


Fig. 3.1 Two daily storage rain gauges in use in the UK: (a) a Met Office Mark II type, and (b) a Snowdon type. Units outside of the parentheses are the original design in inches, while those in parentheses are the equivalent millimetres.

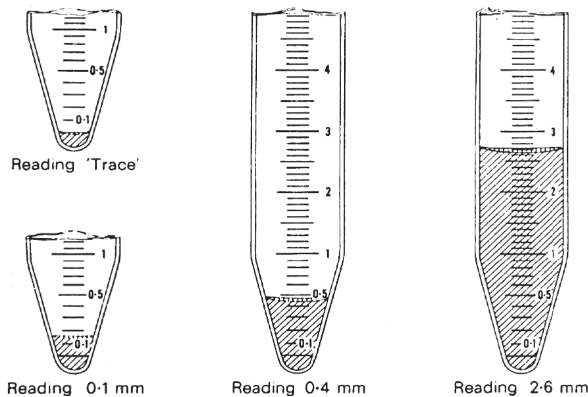


Fig. 3.2 Reading the rain measure. Millimetre graduations. (Adapted from Met Office (1980) *Handbook of Meteorological Instruments*, Vol. 5. Her Majesty's Stationery Office.)

The inner can is easily removed from the outer casing and its contents can be emptied and measured without disturbing the installation.

The gauge is inspected each day at 0900 h GMT, even if it is thought that no precipitation has occurred. Any water in the bottle and inner can is poured into a glass measure (Fig. 3.2) and the reading taken at the lowest point of the meniscus. The glass measure is graduated in relation to the orifice area of the rain gauge and so gives a direct reading of the depth of rain that fell on the area contained by the brass rim. The glass measure has a capacity of 10 mm; if more than 10 mm of rain has fallen, the water in the gauge must be measured in two or more operations. The glass measure is tapered at the bottom so that small quantities can be measured accurately. If no water is found in the gauge and precipitation is known to have fallen, this should be noted as a 'trace' in the records. The glass bottle and inner can should be quite empty before they are returned to the outer case. It is advisable to check the instrument regularly for any signs of external damage, or general wear and tear. Severe frosts can sometimes loosen the joints of the copper casing and, if this is suspected, testing for leaks should be carried out.

The *Snowdon* gauge (Fig. 3.1b), a Met Office Mark I instrument, remains in favour among private observers in the UK, since without the splayed base it is easily maintained in a garden lawn. It is, however, more difficult to keep rigid with the rim level. Globally, the daily storage gauge in most common use is the German *Hellmann* gauge, with over 30 000 gauges of this type in use (Sevruk and Klemm, 1989). This gauge is similar in design to the Snowdon gauge, but with a larger funnel diameter of 159.6 mm.

Monthly rain gauges hold larger quantities of precipitation than daily gauges. The catch is measured using an appropriately graduated glass measure holding 50 mm. Monthly gauges are designed for remote mountain areas and are invaluable on the higher parts of reservoired catchments. Measurements are made on the first day of each month to give the previous month's total and corrections may need to be made

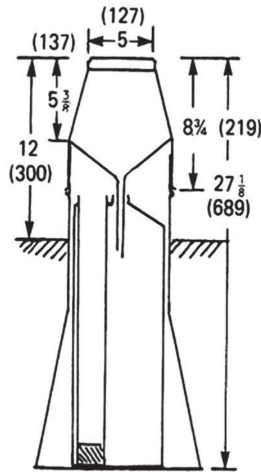


Fig. 3.3 An Octapent monthly storage gauge in use in the UK. Units outside of the parentheses are the original design in inches, while those in parentheses are the equivalent millimetres.

to readings obtained from remote gauges recorded late in the day in wet weather. The *Octapent* monthly rain gauge (Fig. 3.3), a hybrid of the 5-inch and old 8-inch diameter gauges, is made in two sizes with capacities of 685 mm and 1270 mm.

3.2 Recording rain gauges

The need for the continuous recording of precipitation arose from the need to know not just how much rain has fallen, but when it fell and over what period. Numerous instruments have been invented with two main types being widely used: the tilting-siphon rain recorder developed by Dines, and the tipping-bucket gauge, which had its origins with Sir Christopher Wren and Robert Hooke.

The *Dines tilting-syphon* rain recorder (Fig. 3.4) is installed with its rim 500 mm above ground level. The rain falling into the 287 mm diameter funnel is led down to a collecting chamber containing a float. A pen attached to the top of the plastic float marks a chart on a revolving drum driven by clockwork. The collecting chamber is balanced on a knife edge. When there is no rain falling, the pen draws a continuous horizontal line on the chart; during rainfall, the float rises and the pen trace on the chart slopes upwards according to the intensity of the rainfall. When the chamber is full, the pen arm lifts off the top of the chart and the rising float releases a trigger disturbing the balance of the chamber, which tips over and activates the syphon. A counter-weight brings the empty chamber back into the upright position and the pen returns to the bottom of the chart. With double syphon tubes, syphoning should be completed within 8 s, but the rain trap reduces the loss during heavy rainfall. It is recommended, however, that a standard daily storage gauge is installed nearby and that quantities recorded are amended to match the daily total. Each filling of the float chamber is equivalent to 5 mm of precipitation.

Fig. 3.4 The internal mechanism of a tilting-syphon rain gauge. The design is described in Met Office (1982) *Observer's Handbook*, 4th edition. Her Majesty's Stationery Office.

Charts are normally record by the day, but modifications to the instrument can allow a strip chart to be used which gives continuous measurements for as long as a month and which has an extended timescale for intense falls over very short periods. In cold weather, the contents of the float chamber may freeze and special insulation with thermostatically controlled heating equipment, the simplest being a low-wattage bulb, can be installed. The provision of heating assists in the melting of snow, but in very cold weather or during heavy snow, existing low-powered heating devices will not be adequate and there will be a time lag in the melted water being recorded on the chart. Care must be taken to avoid too much heating since evaporation of the melted snow would result in low measurements. Adequate drainage below the gauge should be provided during installation, especially in heavy clay soils and in areas liable to heavy storms, for the syphon system will fail if the delivery pipe enters flood water in the soak away. A model for use in the tropics has a 128-mm diameter receiving aperture and the filling of the float chamber represents 25 mm on the chart. Despite the increased use of tipping-bucket rain gauges with data loggers, tilting-syphon gauges with charts remain in widespread use throughout the tropics.

The principle of the *tipping-bucket* rain gauge is shown in Fig. 3.5. Rain is led down a funnel into a wedge-shaped bucket of fixed capacity. When full, the bucket tips to empty and a twin adjoining bucket begins to fill. At each tip, a magnet attached to the connecting pivot closes a circuit and the ensuing pulse is recorded on a data logger. The mechanism can be used in a variety of gauges. The 15 g of water in one bucketful represents 1 mm of rain caught in a 150 cm^2 gauge, and 0.2 mm in a 750 cm^2 gauge. A small adjustment allows the tipping buckets to be calibrated precisely. It is advisable to install an adjacent storage gauge (sometimes called a 'check gauge') so that a day's or month's total can be measured if the recording mechanism fails. Ceramic resistors connected to a high-capacity battery can be used to reduce the likelihood of the mechanism freezing during cold weather. Other errors specific to tipping-bucket rain gauges are detailed within Hodgkinson *et al.* (2005).