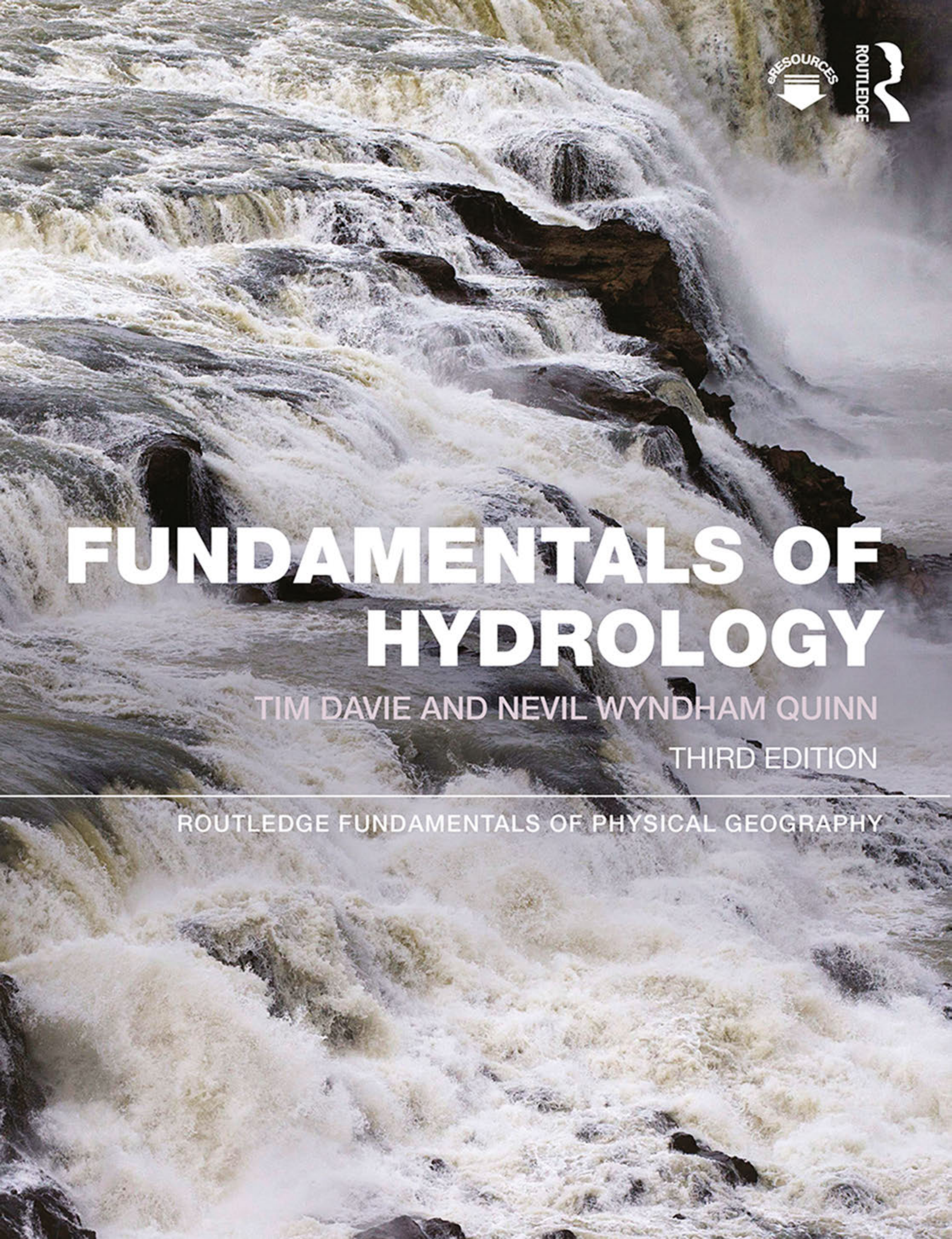

eRESOURCES

ROUTLEDGE

FUNDAMENTALS OF HYDROLOGY

TIM DAVIE AND NEVIL WYNDHAM QUINN

THIRD EDITION

ROUTLEDGE FUNDAMENTALS OF PHYSICAL GEOGRAPHY

FUNDAMENTALS OF HYDROLOGY

The third edition of *Fundamentals of Hydrology* provides an absorbing and comprehensive introduction to the understanding of how fresh water moves on and around the planet and how humans affect and manage the freshwater resources available to them.

The book consists of three parts, each of fundamental importance in the understanding of hydrology:

- The first section deals with processes within the hydrological cycle, our understanding of them, and how to measure and estimate the amount of water within each process. This also includes an analysis of how each process impacts upon water quality issues.
- The second section is concerned with the measurement and analytical assessment of important hydrological parameters such as streamflow and water quality. It describes analytical and modelling techniques used by practising hydrologists in the assessment of water resources.
- The final section of the book draws together the first two parts to discuss the management of freshwater with respect to both water quality and quantity in a changing world.

Fundamentals of Hydrology is a lively and accessible introduction to the study of hydrology at university level. It gives undergraduates a thorough understanding of hydrological processes, knowledge of the techniques used to assess water resources, and an up-to-date overview of water resource management. Throughout the text, examples and case studies from all around the world are used to clearly explain ideas and techniques. Essay questions, guides to further reading, and website links are also included.

Tim Davie is Chief Scientist at the Canterbury Regional Council which controls natural resource management, including water resources, in New Zealand's largest region. Previously he has worked as a research scientist at Landcare Research NZ Ltd and as a university lecturer in Environmental Science and Geography at Queen Mary College, University of London. His PhD, from the University of Bristol, UK, focused on modelling the effects of afforestation on water resources and since then his research interests have been in the fields of land use hydrology (including water quality), satellite remote sensing in hydrology, integrated catchment management, and linkages between science and water policy. Dr Davie is a past president of the New Zealand Hydrological Society and regularly gives guest lectures on water resource management in Universities and to public bodies.

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FUNDAMENTALS OF HYDROLOGY

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SERIES EDITOR'S PREFACE

We are presently living in a time of unparalleled change, and concern for the environment has never been greater. Global warming and climate change, possible rising sea levels, deforestation, desertification, and widespread soil erosion are just some of the issues of current concern. Although it is the role of human activity in such issues that is of most concern, this activity affects the operation of the natural processes that occur within the physical environment. Most of these processes and their effects are taught and researched within the academic discipline of physical geography. A knowledge and understanding of physical geography, and all it entails, is vitally important.

It is the aim of this *Fundamentals of Physical Geography Series* to provide, in five volumes, the fundamental nature of the physical processes that act on or just above the surface of the earth. The volumes in the series are *Climatology*, *Geomorphology*, *Biogeography*, *Hydrology* and *Soils*. The topics are treated in sufficient breadth and depth to provide the coverage expected in a *Fundamentals* series. Each volume leads into the topic by outlining the approach adopted. This is important because there may be several ways of approaching individual topics. Although each volume is complete in itself, there are many explicit and implicit references to the topics covered in the other volumes. Thus, the five volumes together provide a comprehensive insight into the totality that is physical geography.

The flexibility provided by separate volumes has been designed to meet the demand created by the variety of courses currently operating in higher education institutions. The advent of modular courses has meant that physical geography is now rarely taught, in its entirety, in an 'all-embracing' course but is generally split into its main components. This is also the case with many Advanced-level syllabuses. Thus students and teachers are being frustrated increasingly by a lack of suitable books and are having to recommend texts of which only a small part might be relevant to their needs. Such texts also tend to lack the detail required. It is the aim of this series to provide individual volumes of sufficient breadth and depth to fulfil new demands. The volumes should also be of use to sixth form teachers where modular syllabuses are also becoming common.

Each volume has been written by higher education teachers with a wealth of experience in all aspects of the topics they cover and a proven ability in presenting information in a lively and interesting way. Each volume provides a comprehensive coverage of the subject matter using clear text divided into easily accessible sections and subsections. Tables, figures and photographs are used where appropriate as well as boxed case studies and summary notes. References to important previous studies and results are included but are used sparingly to avoid overloading the text. Suggestions for further reading are also provided. The main target readership is introductory level undergraduate students of physical geography or environmental science, but there will be much of interest to students from other disciplines and it is also hoped that sixth form teachers will be able to use the information that is provided in each volume.

John Gerrard

PREFACE TO THE THIRD EDITION

It is 17 years since the first edition of *Fundamentals of Hydrology* was published – time enough to reflect on what has changed in hydrology during this time. One very positive change is that hydrology is now much more integrated within environmental science. It is common to hear reference to catchment science or water management rather than straight hydrology which shows an interest in more than just the physics of water transfer; people are interested in how water affects their health, their livelihoods and the natural world around them. This textbook set out to bring together the discipline of hydrology with aspects of water quality, ecology and natural resource management so it is pleasing to see this type of integrated thinking reflected in scientific literature, university teaching and public debate. If *Fundamentals of Hydrology* has helped in any small way to bring about that change then that is a very positive outcome.

A second area of significant change has been in instrumentation, particularly the rise of fast and small electronic circuitry. This means that we can measure environmental variables in a less intrusive, better and faster way; often continuously rather than at a single point in time. Two obvious examples of this are acoustic doppler streamflow measurement where we can measure river velocities throughout the total water column and optical water quality sensors where we can measure nitrate concentrations continuously in a river. These types of measurements improve our understanding of hydrological processes in both space and time but can also be important information for understanding ecological and land management processes, which in turn promotes the type of integrated science referred to above.

One of the challenges of improved measurement techniques is the quantity of data produced and how to make sense of it all. Fortunately, there has been a corresponding rise in computing power and ability to store these data 'mountains'. An exciting development for this is the rise of artificial intelligence and data mining techniques using fuzzy logic or similar. These types of techniques offer the possibility of making sense of and seeing patterns within enormous data sets, something that was far beyond the capability of hydrologists 20 years ago.

This third edition of *Fundamentals of Hydrology* has been greatly enhanced by the addition of Nevil Quinn as a co-author: Nevil's skills in flood hydrology, water management and up-to-date university teaching has brought a fresh perspective to the text. I am very grateful for his willingness to take on this task and the many hours spent revising and adding new text. I am grateful to the editors at Routledge, Egle Zigaite and Andrew Mould, who have waited patiently for this third edition to be finished. And finally, I am once again thankful to my wife Chris for putting up with disrupted evenings and weekends while I worked on the text.

Tim Davie,
Christchurch, New Zealand November 2018

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HYDROLOGY AS A SCIENCE

INTRODUCTION

Quite literally, hydrology is 'the science or study of' ('logy' from Latin *logia*) 'water' ('hydro' from Greek *hudor*). However, contemporary hydrology does not study all the properties of water. Modern hydrology is concerned with the distribution of water on the surface of the earth; its movement over and beneath the surface, and through the atmosphere. This wide-ranging definition suggests that all water comes under the remit of a hydrologist, while in reality it is the study of fresh water that is of primary concern. The study of the saline water on earth is carried out in oceanography.

When studying the distribution and movement of water it is inevitable that the role of human interaction with it comes into play. Although human needs for water are not the only motivating force in a desire to understand hydrology, they are probably the strongest. This book attempts to integrate the physical processes of hydrology with an understanding of human interaction with fresh water. The human interaction can take the form of water quantity problems (e.g. over-extraction of **groundwater**) or water quality issues (e.g. disposal of pollutants).

> Water is among the most essential requisites that nature provides to sustain life for plants, animals and humans. The total quantity of fresh water on earth could satisfy all the needs of the human population if it were evenly distributed and accessible.
>
> (Stumm 1986: 201)

Although written around 30 years ago, the views expressed by Stumm are still apt today. The real point of Stumm's statement is that water on earth is not evenly distributed and is not evenly accessible. It is the purpose of hydrology as a pure science to explore these disparities and try to explain them. It is the aim of hydrology as an applied science to take the knowledge of why any disparities exist and try to lessen the impact of them. There is much more to hydrology than just supplying water for human needs (e.g. studying floods as natural hazards; the investigation of lakes and rivers for ecological habitats), but analysis of this quotation gives good grounds for looking at different approaches to the study of hydrology.

The two main pathways to the study of hydrology come from engineering and geography, particularly the earth science side of geography. The earth science approach comes from the study of landforms (**geomorphology**) and is rooted in a history of explaining the processes that lead to water moving around the earth and to try to understand spatial links between the processes. The engineering approach tends to be a little more practically based and looks towards finding solutions to problems posed by water moving (or not moving) around the earth. In reality there are huge areas of overlap between the two and it is often difficult to separate them, particularly when you enter into hydrological research. At an undergraduate level, however, the difference manifests itself through earth science hydrology being more descriptive (understanding processes) and engineering hydrology being more numerate (quantifying flows). Within the broad discipline of hydrology there are also areas of specialisation. For example, some hydrologists focus on groundwater and this specialised area is known as geohydrology or hydrogeology. In recent decades another area of specialisation has emerged; that of ecohydrology or hydroecology. This is the study of hydrology in relation to the natural aquatic environment (e.g. rivers and wetlands) and the important interdependence of water and ecosystems.

The approach taken in this book is more towards the earth science side, a reflection of the authors' training and interests, but it is inevitable that there is considerable crossover. There are parts of the book that describe numerical techniques of fundamental importance to any practising hydrologist from whatever background, and it is hoped that the book can be used by all undergraduate students of hydrology.

Throughout the book there are highlighted case studies to illustrate different points made in the text. The case studies are drawn from research projects or different hydrological events around the world and are aimed at reinforcing the text elsewhere in the same chapter. Where appropriate, there are highlighted worked examples illustrating the use of a particular technique on a real data set.

IMPORTANCE OF WATER

Water is the most common substance on the surface of the earth, with the oceans covering over 70 per cent of the planet. Water is one of the few substances that can be found in all three states (i.e. gas, liquid and solid) within the earth's climatic range. The very presence of water in all three forms makes it possible for the earth to have a climate that is habitable for life forms: water acts as a *climate ameliorator* through the energy absorbed and released during transformation between the different phases. In addition to lessening climatic extremes the transformation of water between gas, liquid and solid phases is vital for the transfer of energy around the globe: moving energy from the equatorial regions towards the poles. The low viscosity of water makes it an extremely efficient transport agent, whether through international shipping or river and canal navigation. These characteristics can be described as the *physical properties* of water and they are critical for human survival on planet earth.

The *chemical properties* of water are equally important for our everyday existence. Water is one of the best solvents naturally occurring on the planet. This makes water vital for cleanliness: we use it for washing but also for the disposal of pollutants. The solvent properties of water allow the uptake of vital nutrients from the soil and into plants; this then allows the transfer of the nutrients within a plant's structure. The ability of water to dissolve gases such as oxygen allows life to be sustained within bodies of water such as rivers, lakes and oceans.

The capability of water to support life goes beyond bodies of water; the human body is composed of around 60 per cent water. The majority of this water is within cells, but there is a significant proportion (around 34 per cent) that moves around the body carrying dissolved chemicals which are vital for sustaining our lives (Ross and Wilson 1981). Our bodies can store up energy reserves that allow us to survive without food for weeks but not more than days without water.

There are many other ways that water affects our very being. In places such as Norway, parts of the USA and New Zealand, energy generation for domestic and industrial consumption is through hydro-electric schemes, harnessing the combination of water and gravity in a (by and large) sustainable manner. Water plays a large part in the spiritual lives of millions of people. In Christianity, baptism with water is a powerful symbol of cleansing and God offers 'streams of living water' to those who believe (John 7:38). In Islam there is washing with water before entering a mosque for prayer. In Hinduism, bathing in the sacred Ganges provides a religious cleansing. Many other religions give water an important role in sacred texts and rituals.

Water is important because it underpins our very existence: it is part of our physical, material and spiritual lives. The study of water would therefore also seem to underpin our very existence. Before expanding further on the study of hydrology it is first necessary to step back and take a closer look at the properties of water briefly outlined above. Even though water is the most common substance found on the earth's surface, it is also one of the strangest. Many of these strange properties help to contribute to its importance in sustaining life on earth.

Physical and chemical properties of water

A water molecule consists of two hydrogen atoms bonded to a single oxygen atom (Figure 1.1). The connection between the atoms is through **covalent bonding**: the sharing of an electron from each atom to give a stable pair. This is the strongest type of bonding within molecules and is the reason why water is such a robust compound (i.e. it does not break down into hydrogen and oxygen easily). The robustness of the water molecule means that it stays as a water molecule within our atmosphere because there is not enough energy available to break the covalent bonds and create separate oxygen and hydrogen molecules.

Figure 1.1 shows us that the hydrogen atoms are not arranged around the oxygen atom in a straight line. There is an angle of approximately 105° (i.e. a little larger than a right angle) between the hydrogen atoms. The hydrogen atoms have a positive charge, which means that they repulse each other, but at the same time there are two non-bonding electron pairs on the oxygen atom that also repulse the hydrogen atoms. This leads to the molecular structure shown in Figure 1.1. A water molecule can be described as *bipolar*, which means that there is a positive and negative side to the molecule. This polarity is an important property of water as it leads to the bonding between molecules of water: **hydrogen bonding**. The positive side of the molecule (i.e. the hydrogen side) is attracted to the negative side (i.e. the oxygen atom) of another molecule and a weak hydrogen bond is formed (Figure 1.2). The weakness of this bond means that it can be broken

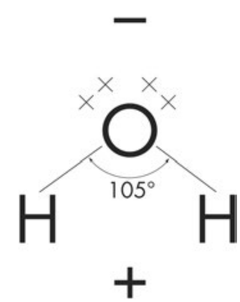

Figure 1.1 The atomic structure of a water molecule. The spare electron pairs on an oxygen atom are shown as small crosses.

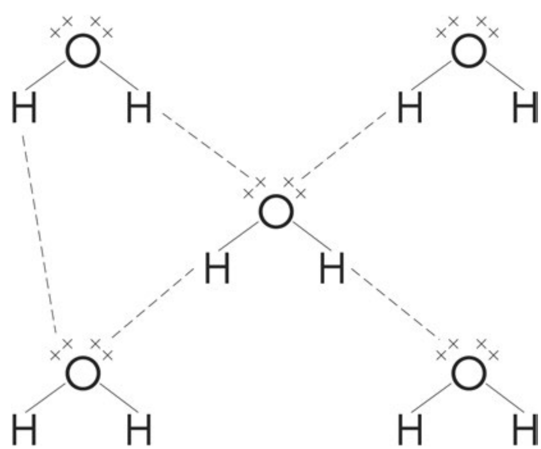

Figure 1.2 The arrangement of water molecules with hydrogen bonds. The stronger covalent bonds between hydrogen and water atoms are shown as solid lines.

Source: Redrawn from McDonald and Kay (1988) and Russell (1976) with the application of some force and the water molecules separate, forming water in a gaseous state (**water vapour**). Although this sounds easy, it actually takes a lot of energy to break the hydrogen bonds between water molecules. This leads to a high specific heat capacity whereby a large amount of energy is absorbed by the water to cause a small rise in energy.

The lack of rigidity in the hydrogen bonds between liquid water molecules gives it two more important properties: a low viscosity and the ability to act as an effective solvent. Low viscosity comes from water molecules not being so tightly bound together that they cannot separate when a force is applied to them. This makes water an extremely efficient transport mechanism. When a ship applies force to the water molecules they move aside to let it pass! The ability to act as an efficient solvent comes through water molecules disassociating from each other and being able to surround charged compounds contained within them. As described earlier, the ability of water to act as an efficient solvent allows us to use it for washing and the disposal of pollutants, and also allows nutrients to pass from the soil to a plant.

In water's solid state (i.e. ice) the hydrogen bonds become rigid and a three-dimensional crystalline structure forms. An unusual property of water is that the solid form has a lower density than the liquid form, something that is rare in other compounds. This property has profound implications for the world we live in as it means that ice floats on water. More importantly for aquatic life, it means that water freezes from the top down rather than the other way around. If water froze from the bottom up, then aquatic flora and fauna would be forced upwards as the water froze and eventually end up stranded on the surface of a pond, river or sea. As it is, the flora and fauna are able to survive underneath the ice in liquid water. The maximum density of water actually occurs at around 4 °C (see Figure 1.3) so that still bodies of water such as lakes and ponds will display thermal stratification, with water close to 4 °C sinking to the bottom.

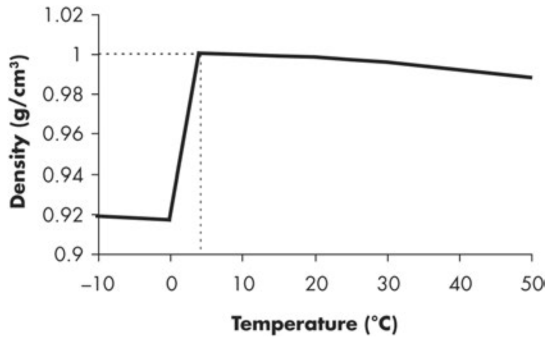

Figure 1.3 The density of water with temperature. The broken line shows the maximum density of water at 3.98 °C.

Water requires a large amount of energy to heat it up. This can be assessed through the **specific heat capacity**, which is the amount of energy required to raise the temperature of a substance by a single degree. Water has a high specific heat capacity relative to other substances (Table 1.1). It requires 4,200 joules of energy to raise the temperature of 1 kilogram of liquid water (approximately 1 litre) by a single degree. In contrast dry soil has a specific heat capacity of around 1.1 kJ/kg/K (it varies according to mineral make up and organic content) and alcohol 0.7 kJ/kg/K. Heating causes the movement of water molecules and that movement requires the breaking of the hydrogen bonds linking them. The large amount of energy required to break the hydrogen bonds in water gives it such a high specific heat capacity.

We can see evidence of water's high specific heat capacity in bathing waters away from the tropics. It is common for sea temperatures to be much lower than air temperatures in high summer since the water is absorbing all the solar radiation and heating up very slowly. In contrast the water temperature also decreases slowly, leading to the sea often being warmer than the air during autumn and winter. As the water cools down it starts to release the energy that it absorbed as it heated up. Consequently for every drop in temperature of 1 °C a single kilogram of water releases 4.2 kJ of energy

Table 1.1 Specific heat capacity of various substances

Substance	*Specific heat capacity (kJ/kg/K)*
Water	4.2
Dry soil	1.1
Ethanol (alcohol)	0.7
Iron	0.44

into the atmosphere. It is this that makes water a climate ameliorator. During the summer months a water body will absorb large amounts of energy as it slowly warms up; in an area without a water body, that energy would heat the earth much quicker (i.e. dry soil in Table 1.1) and consequently air temperatures would be higher. In the winter the energy is slowly released from the water as it cools down and is available for heating the atmosphere nearby. This is why a maritime climate has cooler summers, but warmer winters, than a continental climate.

The energy required to break hydrogen bonds is also the mechanism by which large amounts of energy are transported away from the hot equatorial regions towards the cooler poles. As water evaporates, the hydrogen bonds between liquid molecules are broken. This requires a large amount of energy. The first law of thermodynamics states that energy cannot be destroyed, only transformed into another form. In this case the energy absorbed by the water particles while breaking the hydrogen bonds is transformed into latent heat that is then released as sensible heat as the water precipitates (i.e. returns to a liquid form). In the meantime the water has often moved considerable distances in weather systems, taking the latent energy with it. It is estimated that water movement accounts for 70 per cent of lateral global energy transport through latent heat transfer (Mauser and Schädlich 1998), also known as **advective energy**.

Water acts as a climate ameliorator in one other way: water vapour is a powerful greenhouse gas. Radiation direct from the sun (short-wave radiation) passes straight through the atmosphere and may be then absorbed by the earth's surface. This energy is normally re-radiated back from the earth's surface in a different form (long-wave radiation). The long-wave radiation is absorbed by the gaseous water molecules and consequently does not escape the atmosphere. This leads to the gradual warming of the earth–atmosphere system as there is an imbalance between the incoming and outgoing radiation. It is the presence of water vapour in our atmosphere (and other gases such as carbon dioxide and methane) that has allowed the planet to be warm enough to support all of the present life forms that exist.

Figure 1.4 shows the phase transitions of water and the name of the corresponding process. While some of these processes have already been mentioned, it is important to be familiar with all of them. One that is particularly relevant for the next chapter is **desublimation** or **deposition**. This is where ice forms directly from water vapour. It is also important to note that at normal atmospheric pressure and at temperatures between 0 °C and 100 °C, liquid water is in a stable state, as is water vapour above temperatures of 100 °C, and ice below 0 °C. However water can also exist in metastable states, and importantly these often occur in the atmosphere. Between temperatures of 0 °C and as low as −40 °C, metastable water can exist in liquid form, known as **supercooled water**. Equally, metastable water vapour can exist alongside stable ice and metastable supercooled water. When supercooled liquid water comes into contact with ice, instantaneous freezing occurs. Note that some meteorologists use sublimation to mean both a phase transformation from solid to gas, and also the reverse process. To avoid confusion we will use the equivalent terms deposition and desublimation to refer to the process of a gas becoming a solid without the intermediate liquid phase.

The catchment or river basin

In studying hydrology the most common spatial unit of consideration is the **catchment** or

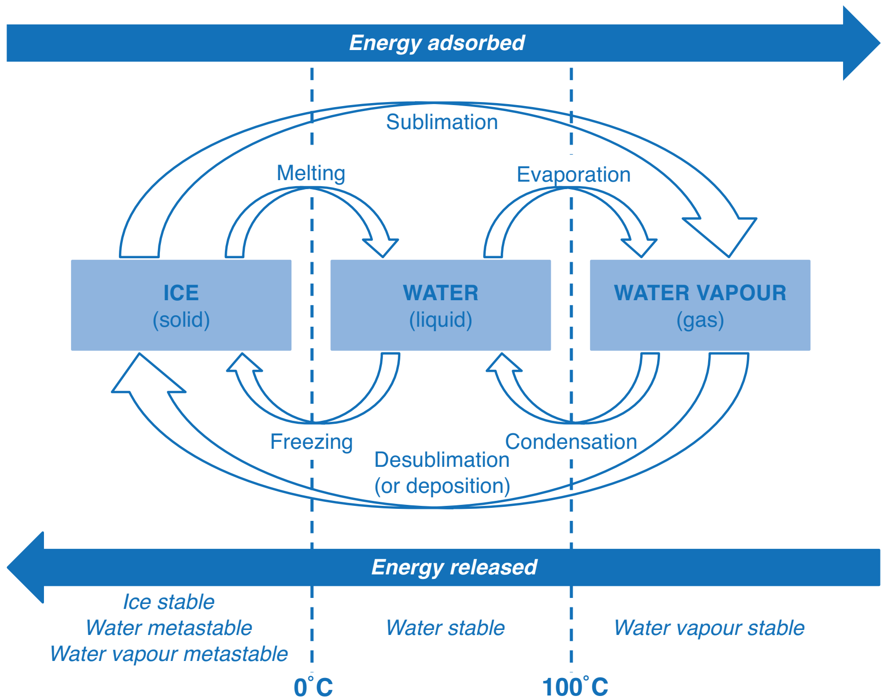

Figure 1.4 Phase changes of water under normal atmospheric conditions and related terminology.

Source: Adapted from Kump et al. (2011)

river basin. This can be defined as the area of land from which water flows towards a river and then in that river to the sea. The terminology suggests that the area is analogous to a basin where all water moves towards a central point (i.e. the plug hole of a basin, or in this case, the river mouth). The common denominator of any point in a catchment is that wherever rain falls, it will end up in the same place: where the river meets the sea (unless lost through evaporation). A catchment may range in size from a matter of hectares to millions of square kilometres, and all catchments are, in reality, made up of a set of nested sub-catchments.

A river basin can be defined in terms of its topography through the assumption that all water falling on the surface flows downhill. In this way a catchment boundary (or divide) can be drawn (catchment delineation) (as in Figure 1.5) which defines the actual catchment area for a river basin. In some parts of the world a river basin is also referred to as a **watershed** – this word stems from the fact that at the catchment boundary water is either 'shed' into one basin or an adjacent basin. Strictly speaking therefore 'watershed' refers to the catchment boundary or divide. The assumption that all water flows downhill to the river is not always correct, especially where the underlying geology of a catchment is complicated. It is possible for water to flow as groundwater into another catchment area, creating a problem for the definition of 'catchment area'. This means that the surface water catchment and the groundwater catchment are not necessarily the same (Figure 1.6). These

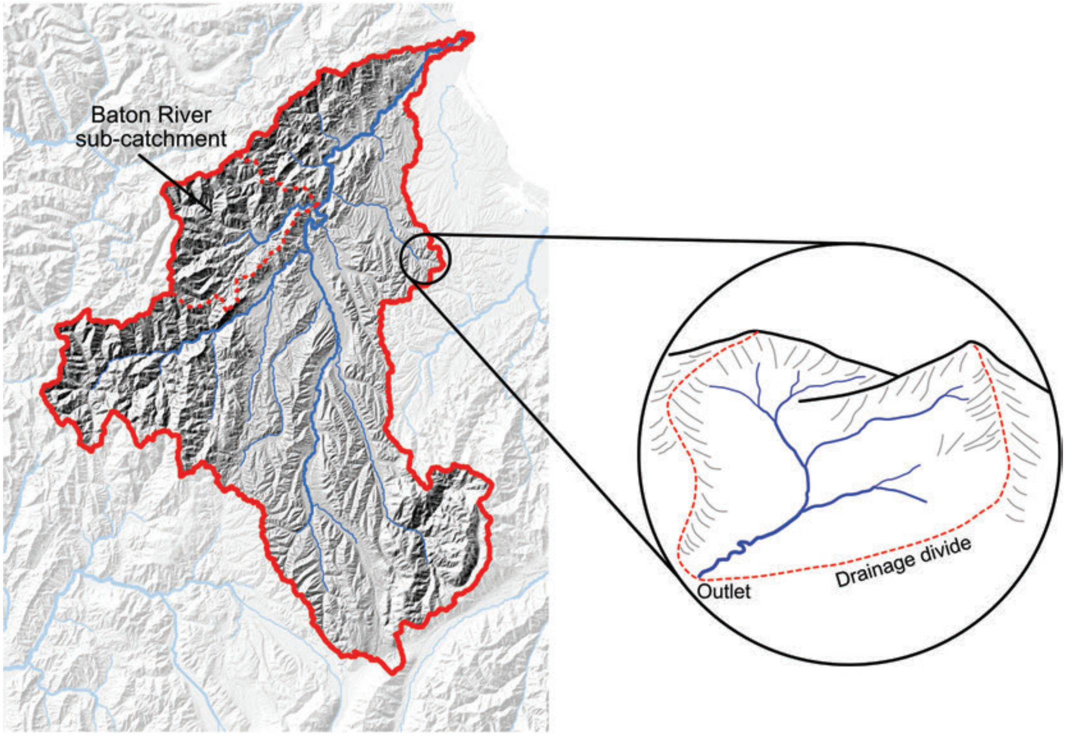

Figure 1.5 Left: Map of the Motueka catchment/watershed, a 2,180 km^2 catchment draining northward at the top of the South Island, New Zealand. Topography is indicated by shading. The Baton river sub-catchment is represented by the dotted outline. Right: A schematic view of a typical small sub-catchment.

Source: Digital elevation model based on USGS 2006 Shuttle Radar Topography Mission. Catchment schematic from Charlton (2008)

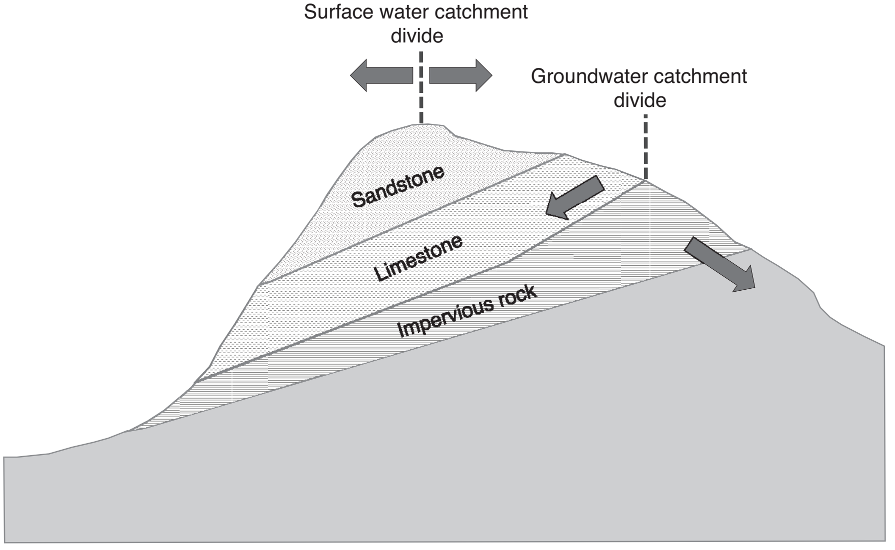

Figure 1.6 The difference between a surface water divide and a groundwater divide. Arrows represent the direction of surface and groundwater flow.

problems aside, the catchment does provide an important spatial unit for hydrologists to consider how water is moving about and is distributed at a certain time.

THE HYDROLOGICAL CYCLE

As a starting point for the study of hydrology it is useful to consider the **hydrological cycle**. This is a conceptual model of how water moves around between the earth and atmosphere in different states as a gas, liquid or solid. As with any conceptual model it contains many gross simplifications; these are discussed in this section. There are different scales at which the hydrological cycle can be viewed, but it is helpful to start at the large global scale and then move to the smaller hydrological unit of a river basin or catchment.

The global hydrological cycle

Table 1.2 sets out an estimate for the amount of water held on the earth at a single time. These figures are extremely hard to estimate accurately. Estimates cited in Gleick (1993) show a range in total from 1.36 to 1.45 thousand million (or US billion) cubic kilometres of water. The vast majority of this is contained in the oceans and seas. If you were to count groundwater less than 1km in depth as 'available' and discount snow and ice, then the total percentage of water available for human consumption is around 0.27 per cent. Although this sounds very little it works out at about 146 million litres of water per person per day (assuming a world population of 7 billion); hence the ease with which Stumm (1986) was able to state that there is enough to satisfy all human needs.

Figure 1.7 shows the movement of water around the earth–atmosphere system and is a representation of the global hydrological cycle. The cycle consists of **evaporation** of liquid water into water vapour that is moved around the atmosphere. At some stage the water vapour condenses into a liquid (or solid) again and falls to the surface as **precipitation**. The oceans evaporate more water than they receive as precipitation, while the opposite is true over the continents. The difference between precipitation and evaporation in the terrestrial zone is **runoff**, water moving over or under the surface towards the oceans, which completes the hydrological cycle. As can be seen

Table 1.2 Estimated volumes of water held at the earth's surface

	Volume ($\times 10^3$ km^3)	*Percentage of total*
Oceans and seas	1,338,000	96.54
Ice caps and glaciers	24,064	1.74
Groundwater	23,400	1.69
Permafrost	300	0.022
Lakes	176	0.013
Soil	16.5	0.001
Atmosphere	12.9	0.0009
Marsh/wetlands	11.5	0.0008
Rivers	2.12	0.00015
Biota	1.12	0.00008
Total	1,385,984	100.00

Source: Data from Shiklomanov and Sokolov (1983)

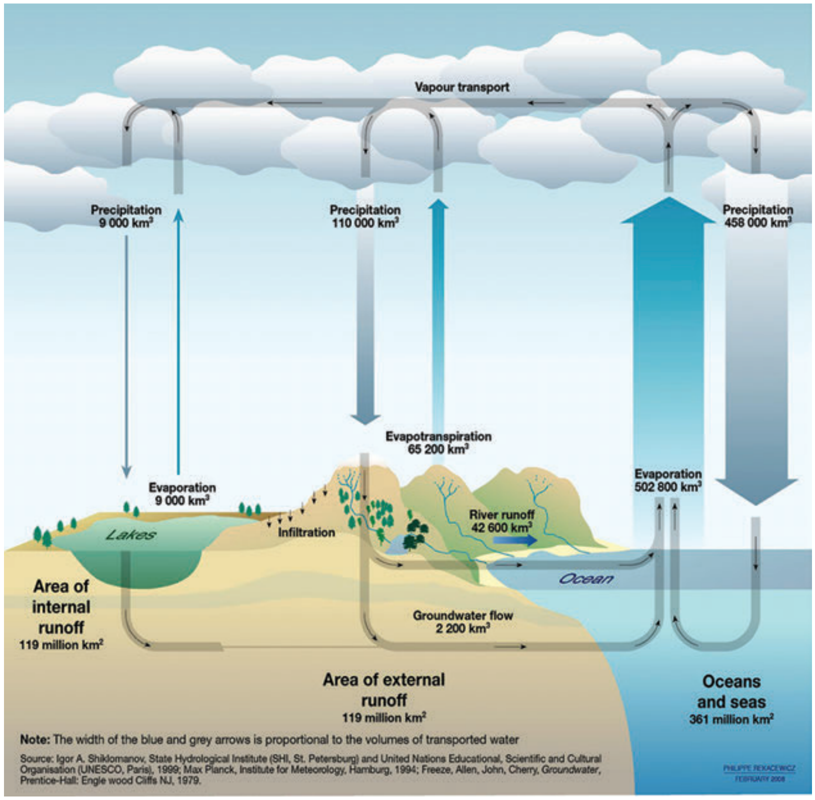

Figure 1.7 The global hydrological cycle. The numbers represent estimates on the total amount of water (km^3) in each process per annum. The thickness of the arrows denotes the proportional volume.

Source: Figure drawn by Philippe Rekacewicz (GRID-Arendal) (based on data from UNEP (2008))

in Figure 1.7, where the width of the arrows is proportional to the volume, the vast majority of evaporation and precipitation occurs over the oceans. Ironically this means that the terrestrial zone, which is of greatest concern to hydrologists, is actually rather insignificant in global terms.

The processes shown in Figure 1.7 (evaporation, precipitation and runoff) are the fundamental processes of concern in hydrology. The figures given in the diagram are global totals, but they vary enormously around the globe. This is illustrated in Figure 1.8 which shows how total precipitation is partitioned towards different hydrological processes in differing amounts depending on climate. In temperate climates (i.e. non-tropical or polar) around one third of precipitation becomes evaporation, one third surface runoff and the final third as groundwater recharge. In arid and semi-arid regions the proportion of evaporation is much greater, at the expense of groundwater recharge.

With the advent of satellite monitoring of the earth's surface in the past 40 years it is now possible to gather information on the global distribution

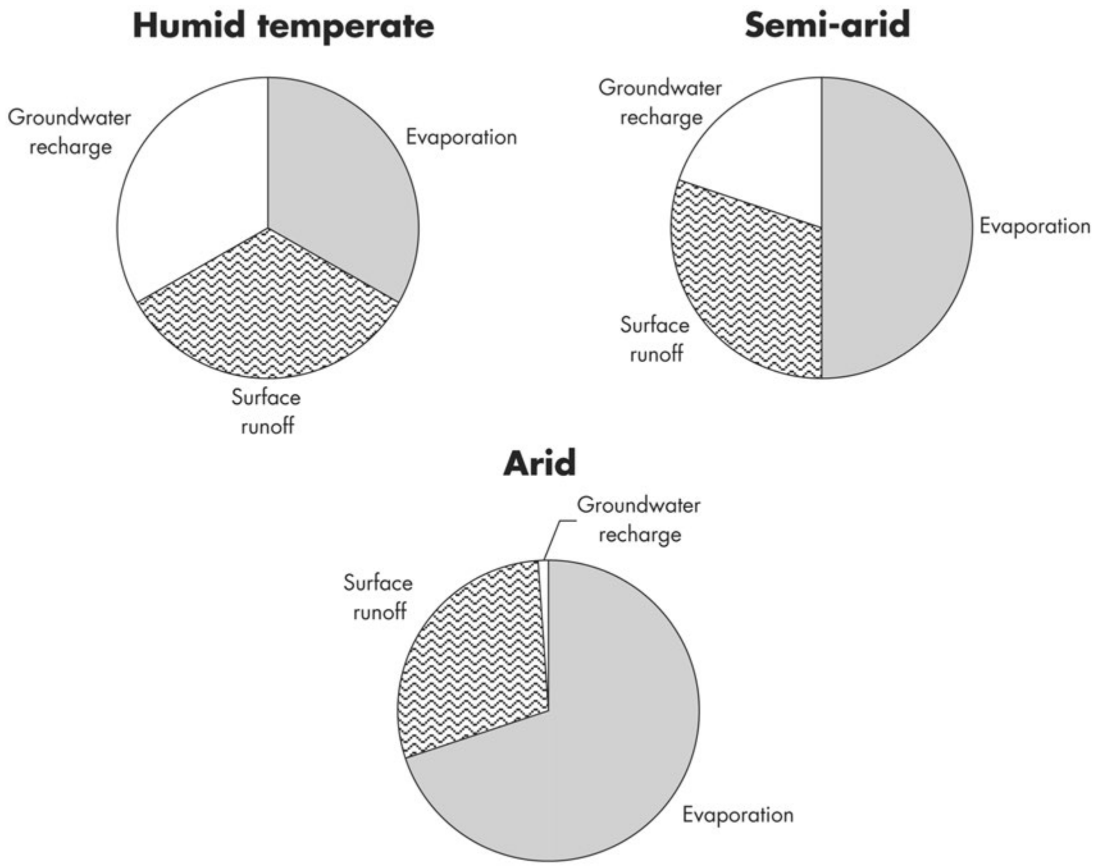

Figure 1.8 Proportion of total precipitation that returns to evaporation, surface runoff or groundwater recharge in three different climate zones.

Source: UNESCO (2006)

of these three processes and hence view how the hydrological cycle varies around the world.

The global distribution of freshwater resources

The figure given above of 146 million litres of fresh water per person per year is extremely misleading, as the distribution of available water around the globe varies enormously. The concept of available water considers not only the distribution of rainfall but also population. Table 1.3 gives some indication of those countries that could be considered water-rich and water-poor in terms of available water. Even this is misleading as a country such as Australia is so large that the high rainfall received in the tropical north-west compensates for the extreme lack of rainfall elsewhere; hence it is considered water-rich. The use of rainfall alone is also misleading as it does not consider the importation of water passing across borders, through rivers and groundwater movement.

Table 1.3 gives the amount of available water for various countries, but this takes no account for

Table 1.3 Annual renewable water resources per capita (2013 figures) of the seven water resource-richest and poorest countries (and other selected countries). Annual renewable water resource is based upon the rainfall within each country; in many cases this is based on estimated figures

Water resource richest countries	*Annual internal renewable water resources per capita (thousand m³/yr)*	*Water resource poorest countries*	*Annual internal renewable water resources per capita (thousand m³/yr)*
Iceland	525.074	Kuwait	0.000
Guyana	301.396	Bahrain	0.003
Suriname	183.579	United Arab Emirates	0.016
Papua New Guinea	109.407	Egypt, Arab Rep.	0.022
Bhutan	103.456	Qatar	0.026
Gabon	98.103	Bahamas, The	0.053
Canada	81.071	Sudan	0.081
Other selected countries			
Australia	21.272	South Africa	0.843
USA	8.914	Kenya	0.467
United Kingdom	2.262	Israel	0.093

Source: Data from World Bank Indicators 2014 (http://data.worldbank.org/indicator/ER.H2O.INTR.PC)

the amount of water abstracted for actual usage. Figure 1.8 shows the water abstraction per capita for all of the OECD countries. This shows that the USA, Canada and Australia are very high water-users, reflecting a very large amount of water used for agricultural and industrial production. The largest water user is the USA with 1,730m^3 per capita per annum, which is still only 1 per cent of the 146 million litres per capita per annum derived from the Stumm quote. Australia as a high water-user has run into enormous difficulties in the years 2005–2007 with severe drought, limiting water availability for domestic and agricultural users. In a situation like this the way that water is allocated (see Chapter 11) literally becomes a matter of life and death, and many economic livelihoods depend on equitable allocation of a scarce water resource.

To try and overcome some of the difficulties in interpreting the data in Figure 1.7 and Table 1.2, hydrologists often work at a scale of more relevance to the physical processes occurring. This is frequently the water basin or catchment scale (Figure 1.5).

The catchment hydrological cycle

At a smaller scale it is possible to view the catchment hydrological cycle as a more in-depth conceptual model of the hydrological processes operating. Figure 1.10 shows an adaptation of the global hydrological cycle to show the processes operating within a catchment. In Figure 1.10 there are still essentially three processes operating (evaporation, precipitation and runoff), but it is possible to subdivide each into different sub-processes. Evaporation is a mixture of open water evaporation (i.e. from rivers and lakes); evaporation from the soil; evaporation from plant surfaces; **interception**; and **transpiration** from plants. Precipitation can be in the form of **snowfall**, hail, rainfall or some mixture of the three (sleet). Interception of precipitation by plants makes the water available for evaporation again before it even reaches the soil surface. Buildings and impervious surfaces similarly intercept precipitation and what proportion is evaporated or becomes runoff. The broad term ‘runoff’ incorporates the movement of liquid water above and

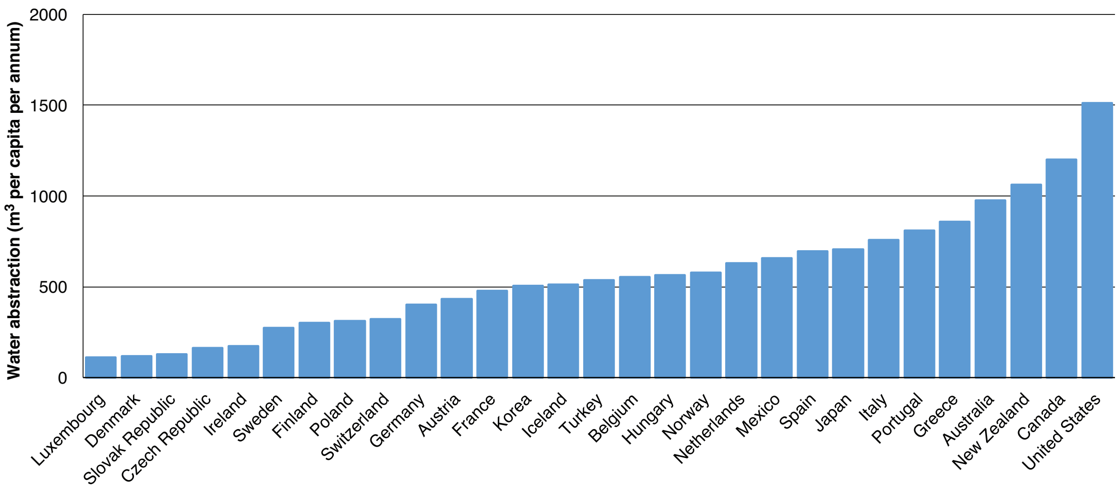

Figure 1.9 Water abstracted per capita for the OECD countries.

Source: World Bank Indicators (2014)

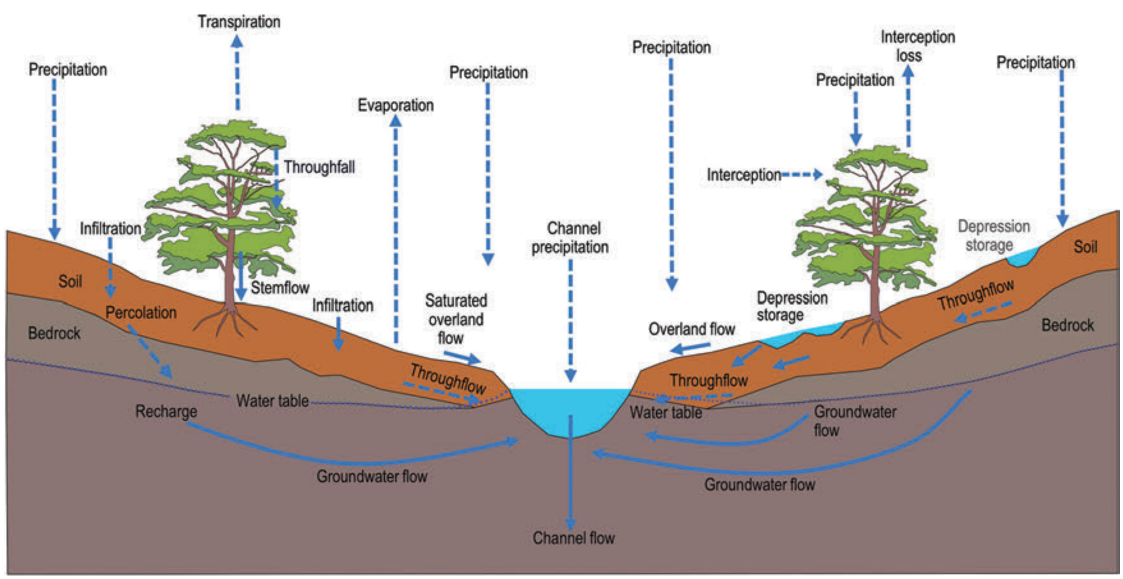

Figure 1.10 Processes in the hydrological cycle operating at the basin or catchment scale.

Source: Adapted from Bishop and Prosser (2001) below the surface of the earth. The movement of water below the surface necessitates an understanding of infiltration into the soil and how the water moves in the unsaturated zone (**throughflow**) and in the saturated zone (**groundwater flow**). All of these processes and sub-processes are dealt with in detail in later chapters; what is important to realise at this stage is that it is part of one continuous cycle that moves water around the globe and that they may all be operating at different times within a river basin.

THE WATER BALANCE EQUATION

In the previous section it was stated that the hydrological cycle is a conceptual model representing our understanding of which processes are operating within an overall earth–atmosphere system. It is also possible to represent this in the form of an equation, which is normally termed the **water balance equation**. The water balance equation is a mathematical description of the hydrological processes operating within a given timeframe and incorporates principles of mass and energy continuity. In this way the hydrological cycle is defined as a closed system whereby there is no mass or energy created or lost within it. The mass of concern in this case is water.

There are numerous ways of representing the water balance equation, but Equation 1.1 shows it in its most fundamental form.

$$P \pm E \pm \Delta S \pm Q = 0 \quad (1.1)$$

where P is precipitation; E is evaporation; ΔS is the change in **storage** and Q is runoff. Runoff is normally given the notation of Q to distinguish it from rainfall which is often given the symbol R and frequently forms the major component of precipitation. The ± terminology in Equation 1.1 represents the fact that each term can be either positive or negative depending on which way you view it – for example, precipitation is a gain (positive) to the earth but a loss (negative) to the atmosphere. As most hydrology is concerned with water on or about the earth's surface it is customary to consider the terms as positive when they represent a gain to the earth.

If we think about the water balance equation from the perspective of trying to quantify flow in a river (Q) then we can re-arrange it as shown in Equation 1.2.

$$Q = P - E - \Delta S \quad (1.2)$$

In Equations 1.1 and 1.2 the change in storage term can be either positive or negative, as water can be released from storage (negative) or absorbed into storage (positive).

The terms in the water balance equation can be recognised as a series of fluxes and stores. A **flux** is a rate of flow of some quantity (Goudie et al. 1994): in the case of hydrology the quantity is water. The water balance equation assesses the relative flux of water to and from the surface with a storage term also incorporated. A large part of hydrology is involved in measuring or estimating the amount of water involved in this flux transfer and storage of water.

Precipitation in the water balance equation represents the main input of water to a surface (e.g. a catchment). As explained on p. 11, precipitation is a flux of both rainfall and snowfall. Evaporation as a flux includes that from open water bodies (lakes, ponds, rivers), the soil surface and vegetation (including both interception and transpiration from plants). The storage term includes soil moisture, deep groundwater, water in lakes, glaciers, and seasonal snow cover. The runoff flux is also explained on p. 11. In essence it is the movement of liquid water above and below the surface of the earth.

The water balance equation is probably the closest that hydrology comes to having a fundamental theory underlying it as a science, and hence almost all hydrological study is based around it. Field catchment studies are frequently trying to measure the different components of the equation in order to assess others. Nearly all hydrological **models** attempt to solve the equation for a given time span – for example, by knowing the amount of rainfall for a given area and estimating the amount of evaporation and change in storage it is possible to calculate the amount of runoff that might be expected.

Despite its position as a fundamental hydrological theory there is still considerable uncertainty about the application of the water balance equation. It is not an uncertainty about the equation itself but rather about how it may be applied. The problem is that all of the processes occur at a spatial and temporal scale (i.e. they operate over a period of time and within a certain area) that may not coincide with the scale at which we make our measurement or estimation. It is this issue of *scale* that makes hydrology appear an imprecise science and it will be discussed further in the remaining chapters of this book. In the next section we will consider an important aspect of temporal scale that will underpin much of the later discussion around understanding and quantifying processes such as rainfall and runoff.

MAGNITUDE–FREQUENCY–DURATION RELATIONSHIPS

All components of the hydrological cycle vary over time. For example, flow in a stream might vary seasonally, but also over the short term (e.g. hours) in response to a rainfall event. Describing and understanding how processes vary temporally is an important aspect of hydrology, and plotting the frequency distribution of a process can reveal important information. For example, in Figure 1.11, the frequency distribution of daily mean river flows

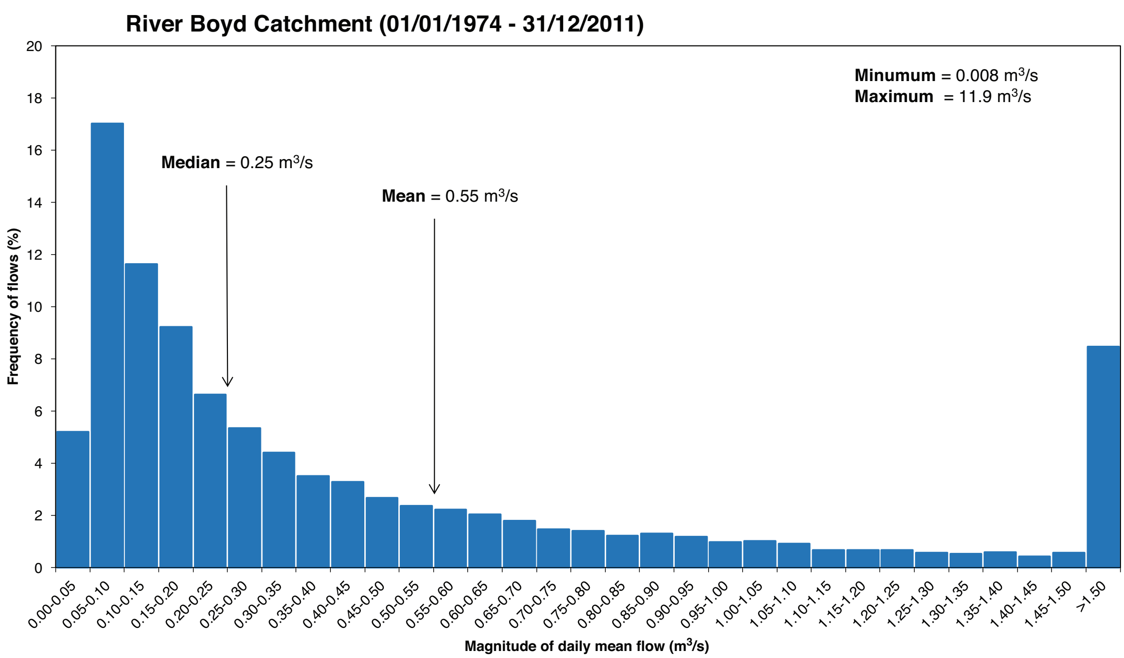

Figure 1.11 Frequency of flows in the River Boyd catchment near Bitton, UK, for the period 1974 to 2011.

Source: Data from the National River Flow Archive, Centre for Ecology and Hydrology, United Kingdom measured at a gauging station over a 38-year period shows that for about half the time, flows are typically in the range of 0.1 to 0.3 m^3/s. Occasionally (about 5 per cent of the time) flows are less than this, and even more occasionally flows are higher, and much higher, than this. The highest flow of nearly 12 m^3/s, only on one day of the 38-year record, and flows exceeding 9 m^3/s occur only on 7 days, or 0.05 per cent of the time. What this illustrates is a fundamental relationship between *magnitude* and *frequency*. High-magnitude events occur very infrequently, in other words, extreme events are rare. The occurrence of high magnitude, low frequency events is also responsible for the asymmetry of the distribution in Figure 1.11. Whereas a normal distribution would have flows symmetrically distributed about the mean, flow (and rainfall) are positively skewed, often with a long tail of higher values. As a result the median is often much less than the mean.

Understanding magnitude–frequency relationships is an important aspect of hydrology, particularly in relation to flood hydrology, and we will cover this in a little more detail in Chapter 9. What is important to remember now, is that regardless of the type of hydrological process (e.g. precipitation, flow), each measurement or estimation of magnitude is associated with a frequency which reflects how commonly or rarely events of that magnitude occur. This means that the frequency also tells us something about the *likelihood* of occurrence. This is also referred to as the *probability* of occurrence and can be expressed in a decimal (or percentage) form where 1 (or 100 per cent) would be certainty of occurrence, and 0 would be certainty of non-occurrence. For example if a weather report suggests an 80 per cent chance of rain, this means that under similar meteorological conditions in the past it has rained 8 out of 10 times. Implicit in this is that probability is calculated from the relative frequency:

$$p = \frac{n}{N} \tag{1.3}$$

where

p = probability
n = number of occurrences
N = total number of observations

As we will see in Chapter 9, the inverse of the probability is the recurrence interval. So for example, if we know that a particular flow has a probability of 1 per cent (p = 0.01), then the recurrence interval is 1/0.01 = 100 years. So the event that has a 1 per cent chance of occurring is also the 1 in 100 year event. In the past this was known as the 'return period' but the term recurrence interval is now preferred because the concept of a 'return period' is frequently misunderstood. For example, if a 1 in 100 year flood was experienced last year this does not mean another similarly sized flood will not occur for another 100 years. In fact, the chances of experiencing a similar flood in the next year remain the same. What the 'return period' or recurrence interval really refers to is the average time between events of the same magnitude – but averaged over a long period.

So, we have seen how magnitude and frequency are related, and how these can tell us something about event rarity and the probability of occurrence. We will now turn to another important descriptor in hydrology, that of *duration*. Fundamental relationships between magnitude and frequency also apply when we consider duration. For example, we can distinguish between the total depth of rain falling in 1 hour, several hours, or a 24-hour period. So for a particular location, we would have different estimates of magnitude (depth) depending on both duration and recurrence interval (frequency) (Figure 1.12).

Magnitude–frequency–duration relationships are characteristic of all the processes we will be covering in this book, but are especially important in relation to rainfall and river flow. The reason for dealing with these fundamentals here is to signal the importance of being aware of these three dimensions and their interrelationships.

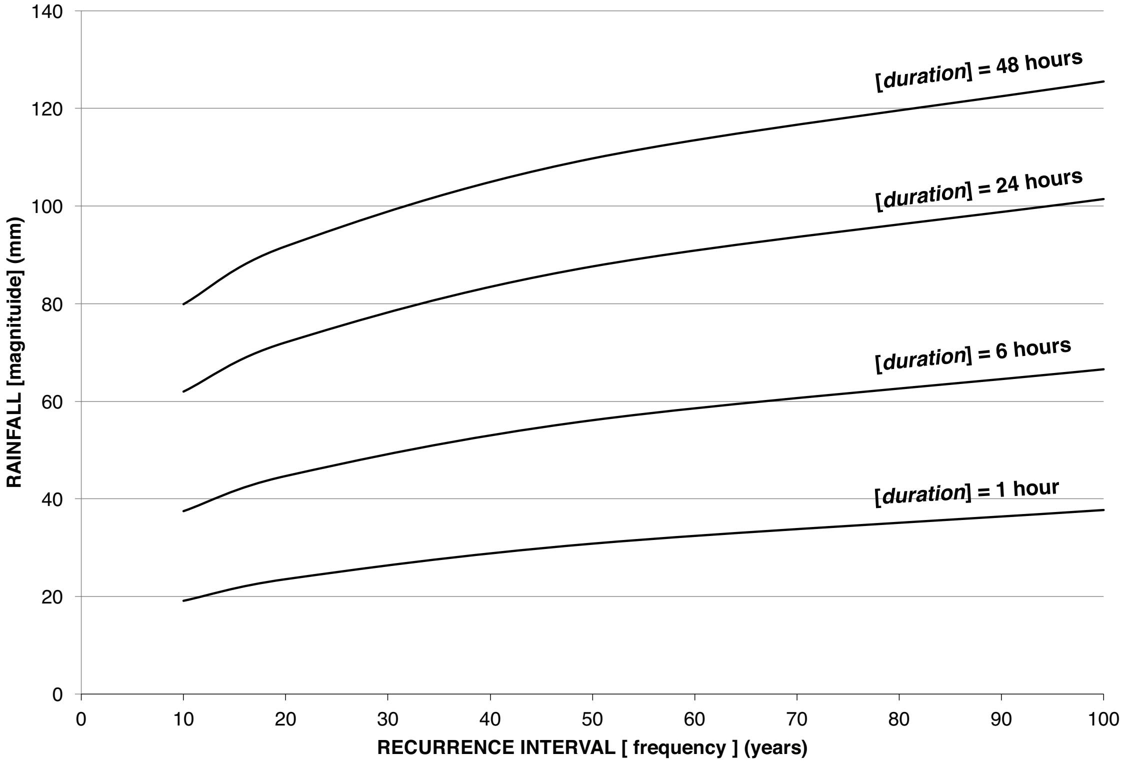

Figure 1.12 Rainfall magnitude–frequency–duration relationships for the River Boyd catchment, United Kingdom.
Source: Data from CEH (1999)

OUTLINE OF THE BOOK

Fundamentals of Hydrology attempts to bring out the underlying principles in the science of hydrology and place these in a water management context. By and large, water management is concerned with issues of water quantity (e.g. floods, droughts, water supply and distribution) and water quality (e.g. drinking water, aquatic ecosystem health). These two management concerns form the basis for discussion within the book. The first section focuses on components of the hydrological cycle. Precipitation is dealt with in Chapter 2, followed by evaporation in Chapter 3, each of these being a key term in the water balance equation. Chapter 4 considers what happens to the water intercepted by vegetation and the land surface, introducing the storage term from the water balance equation, particularly in relation to the role of water stored as snow and ice. Chapters 5 and 6 look at the two other major storages; that under the earth's surface as groundwater (Chapter 5) and that within the soil (Chapter 6). Even though water has to pass through the soil before becoming groundwater, we deal with groundwater first because the fundamental principles governing flow in groundwater (saturated flow) are a little more straightforward than flow in the soil under unsaturated conditions. Chapter 7 is concerned with the runoff processes that lead to water flowing down a channel in a stream or river.

Each of Chapters 2–7 starts with a detailed description of the process under review in the chapter. They then move on to contain a section on how it is possible to measure the process, followed by a section on how it may be estimated. In reality it is not always possible to separate between measurement and estimation as many techniques contain an element of both within them, something that is pointed out in various places within these chapters. Chapters 2–7 finish with a discussion on how the particular process described has relevance to water quantity and quality.

Chapters 8 and 9 move away from descriptions of processes and look at the methods available to firstly measure channel flow (Chapter 8) and then how to analyse these streamflow records (Chapter 9). This is one of the main tasks within hydrology and three particular techniques are described: hydrograph analysis (including the unit hydrograph), flow duration curves and frequency analysis. The latter mostly concentrates on **flood frequency analysis**, although there is a short description of how the techniques can be applied to low flows. The chapter also has sections on hydrological modelling and combining ecology and hydrology for instream flow analysis.

Chapter 10 is concerned with water quality in the fresh water environment. This chapter has a description of major water quality parameters, measurement techniques and some strategies used to control water quality.

The final chapter takes an integrated approach to look at different issues of change that affect hydrology (Chapter 11). These range from water resource management and a changing legislative framework to climate and land use change. These issues are discussed with reference to research studies investigating the different themes. It is intended as a way of capping off the fundamentals of hydrology by looking at real issues facing hydrology in the twenty-first century.

ESSAY QUESTIONS

1. **Discuss the nature of water's physical properties and how important these are in determining the natural climate of the earth.**
2. **Describe how the hydrological cycle varies around the globe.**
3. **How may water-poor countries overcome the lack of water resources within their borders?**
4. **Using examples, explain the concept of a magnitude–frequency–duration relationship and its importance in hydrology.**

WEBSITES

A warning: although it is often easy to access information via the internet you should always be careful in utilising it. There is little control on the type of information available or on the data presented. More traditional channels, such as research journals and books, undergo a peer review process where there is some checking of content. This may happen for websites, but there is no guarantee that it has happened. You should be wary of treating everything read from the internet as being correct.

The websites listed here are general sources of hydrological information that may enhance the reading of this book. The majority of addresses are included for the web links provided within their sites. The web addresses were up to date in early 2019 but may change in the future. Hopefully there is enough information provided to enable the use of a search engine to locate updated addresses.

International associations and national hydrological societies

http://iahs.info
International Association of Hydrological Sciences (IAHS): a constituent body of the International Union of Geodesy and Geophysics (IUGG), promoting the interests of hydrology around the world. This has a useful links page.

www.aihydrology.org
Home page of the American Institute of Hydrology.

www.hydrology.org.uk
Home page of the British Hydrological Society.
http://cgu-ugc.ca/sections/hydrology
Home page of the Hydrology Section of the Canadian Geophysical Union.
www.hydrologynz.org.nz
Home page of the New Zealand Hydrological Society: has a links page with many hydrological links.
www.wmo.int/pages/prog/hwrp/publications/international_glossary/385_IGH_2012.pdf
This is the WMO International Glossary of Hydrology, providing definitions in several languages.
https://or.water.usgs.gov/projs_dir/willgw/glossary.html
This is the United States Geological Survey (USGS) Glossary of Hydrologic Terms.

Hydrological data

www.worldwater.org
The World's Water, part of the Pacific Institute for Studies in Development, Environment, and Security: this is an organisation that studies water resource issues around the world. There are some useful information sets here.
www.usgs.gov/water
Water Resources Division of the USGS: provides information on groundwater, surface water and water quality throughout the USA.
https://ghrc.nsstc.nasa.gov/home
Global Hydrology Resource Centre: a NASA site with mainly remote sensing data sets of relevance for global hydrology.
https://hydrohub.wmo.int/en/world-hydrological-cycle-observing-system-whycos
WHYCOS is a World Meteorological Organization (WMO) programme aiming at improving the basic observation activities, strengthening international cooperation and promoting free exchange of data in the field of hydrology. This website provides information on the System, projects, technical materials, data and links.
www.who.int/water_sanitation_health/en
This World Health Organization (WHO) section contains fact sheets on over 20 water-related diseases, estimates of the global burden of water-related disease, information on water requirements (quantity, service level) to secure health benefits, and facts and figures on water, sanitation and hygiene links to health.
https://en.unesco.org/themes/water-security
UNESCO has an active programme in water security and also hosts the International Hydrological Programme (IHP).

Hydrological research

www.ucowr.org
Universities Council on Water Resources: 'universities and organizations leading in education, research and public service in water resources'. Disseminates information of interest to the water resources community in the USA.
www.wsag.unh.edu
Water Systems Analysis Group at the University of New Hampshire: undertakes a diverse group of hydrological research projects at different scales and regions. Much useful information and many useful links.
www.ceh.ac.uk
Centre for Ecology and Hydrology (formerly Institute of Hydrology) in the UK: a hydrological research institute. There is a very good worldwide links page. CEH also hosts the UK's National River Flow Archive (**https://nrfa.ceh.ac.uk**)
http://water.oregonstate.edu
Hillslope and Watershed Hydrology Team at Oregon State University: this has many good links and information on the latest research.

2

PRECIPITATION

LEARNING OBJECTIVES

When you have finished reading this chapter you should have:

- An understanding of the types of precipitation, the fundamental process of precipitation formation and the different precipitation mechanisms.
- A knowledge of the techniques for measuring precipitation (rainfall and snow).
- An appreciation of the associated errors in measuring precipitation.
- A knowledge of how to analyse rainfall data spatially and for intensity/duration of a storm.
- A knowledge of some of the methods used to estimate rainfall at the large scale.

PRECIPITATION AS A PROCESS

Precipitation is the release of water from the atmosphere to reach the surface of the earth. The term 'precipitation' covers all forms of water being released by the atmosphere, including snow, hail, sleet and rainfall. It is the major input of water to a river catchment area and as such needs careful assessment in any hydrological study. Although rainfall is relatively straightforward to measure (other forms of precipitation are more difficult) it is notoriously difficult to measure *accurately* and, to compound the problem, is also extremely variable within a catchment area.

Understanding the atmosphere

The atmosphere around is comprised of the gases we breathe (principally nitrogen [~78 per cent], but also oxygen [~21 per cent]), and water vapour. Gravity concentrates these gases closer to the earth's surface, which is why atmospheric pressure decreases as you ascend. For example, at 5,500 m, the pressure is half that of the surface, which is why an aircraft cabin needs to be pressurised. The atmospheric pressure measured on the surface by a barometer represents the pressure exerted by all these gases upon the earth's surface. Although the SI unit of pressure is *newtons per square metre* (Nm^{-2}), commonly named the *pascal* (Pa; 1 Pa = 1 Nm^{-2}), pressure is typically reported in millibars (mbar or mb) or hectopascals (hPa) (one standard atmosphere = 1,013.2 mb = 1,013.2 Pa). Water vapour typically accounts for a very small proportion of total air pressure (<0.3 per cent) (Ward and Robinson 2000).

We can simplify the composition of air in the atmosphere into two parts – water vapour, and dry air (being all the other gases), and moist air is therefore the sum of these components. The maximum amount of water vapour that can normally be held in air is a function of temperature and is shown by the curve in Figure 2.1, sometimes called the *Clausius-Claperon* relation. So the **vapour pressure** (often denoted as e) at any particular temperature cannot exceed a maximum (denoted as e_{sat}), known as the **saturated vapour pressure**. The saturation vapour pressure curve represents *absolute humidity* at each temperature, and at this temperature, water vapour starts to condense, forming droplets of water. For this reason it is also known as the *dew point temperature* (or the *frost point temperature* if the temperature is less than 0 °C). Note that in the description above we said the maximum water vapour that can *normally* be held, in some cases this can be exceeded, which would be described as

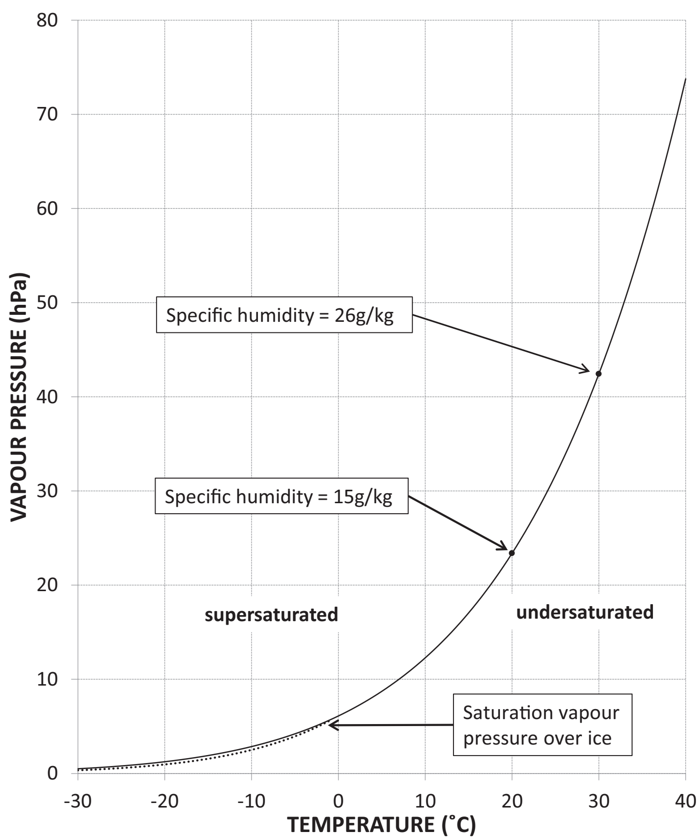

Figure 2.1 Saturation vapour pressure curve representing absolute humidity for a given dew point temperature. Note that the saturation vapour pressure curve over ice is lower.

supersaturated conditions. Superstaturation beyond 1 or 2 per cent in the atmosphere is very rare.

While the solid line in Figure 2.1 represents the saturation vapour pressure over water, the dotted line represents the saturation vapour pressure over ice. The reason for this difference is that it is easier for water molecules to escape from the surface of liquid water than it is from solid water (ice). At temperatures below the frost point temperature, *deposition* (or *desublimation*) occurs where ice forms directly from water vapour without the intermediate phase of liquid water; this is how *hoar frost* forms.

Vapour pressure is a reflection of the concentration of water vapour molecules in a volume of air – effectively a density which is measured as the mass of water vapour molecules (usually g) per unit mass of air (usually kg). Expressing the amount of water vapour in this way is known as the *specific humidity*. Look at the difference between the vapour pressure at 30 °C and 20 °C in Figure 2.1. These vapour pressures correspond to 26 g water vapour per kg of air and 15 g of water vapour per kg of air, respectively. In other words, raising the temperature by 10 °C means that an extra 11 g of water can be absorbed in the air, with a corresponding increase in the vapour pressure. Similarly, if the air was saturated at 30 °C, and the temperature suddenly dropped to 20 °C, 11 g of water vapour would condense as droplets. The latter example explains why droplets of water form on a glass filled with a cold liquid, when it is brought out into the warmth of a summer's day.

The *Ideal Gas Law* (Equation 2.1) tells us that there are fundamental relationships between the pressure and volume of a gas and temperature, and shows that the ratio between pressure on the one hand and the product of temperature and density on the other is constant. This means that if pressure is decreased, then so must be the product of temperature and pressure, in order to retain the same constant:

$$\frac{P}{T_a \cdot \rho_a} = R_a \qquad (2.1)$$

where

P = atmospheric pressure (kPa)
T_a = air temperature (K)
ρ_a = mass density of air (kg/m^3)
R_a = gas constant (0.288)

This means when a parcel of air moves upwards in the atmosphere to an area of lower pressure, it expands (density decreases) and cools. The cooling referred to here has nothing to do with a reduction in heat input (like the effect of switching off a stove plate), it occurs simply because of the change in pressure and density. This is known as *adiabatic cooling* and is fundamentally important in the precipitation formation process. The *dry adiabatic lapse rate* (DALR) quantifies this reduction in temperature as an air mass moves upwards, and is approximately 9.8 °C per km. This would apply if the air was dry, but if it is saturated, then the slower *saturated adiabatic lapse rate* (SALR) applies, because latent heat from the condensation process is released which offsets the cooling process. The SALR can be as much as half the DALR, but its precise value depends on prevailing air pressure and temperature. In addition to the adiabatic lapse rates there is also the *environmental lapse rate* (ELR). This is related to the more conventional understanding of how temperature decreases the further you are away from a radiating body (the earth), and is equivalent to about 6 °C per km, but does vary.

PRECIPITATION FORMATION

As we have seen above, the ability of air to hold water vapour is temperature dependent: the cooler the air the less water vapour is retained. If a body of warm, moist air is cooled then it will become saturated with water vapour and eventually the water vapour will condense into liquid or solid water (i.e. cloud droplets or ice crystals). The water will not condense spontaneously however; there need to be minute particles present in the atmosphere, called **condensation nuclei**, upon which the cloud droplets or ice crystals form. The cloud droplets or ice crystals that form on condensation nuclei are normally too small to fall to the surface as precipitation; they need to grow in order to have enough mass to fall. Their mass acting under gravity results in a downward force, which is also subject to air resistance (or drag forces). When the downward force equals the drag force, the drop reaches terminal velocity. These droplets or ice particles need also overcome any uplifting forces within a cloud in order to fall. Figure 2.2 shows the relative sizes of condensation nuclei, cloud drops and rain drops. So there are three conditions that need to be met prior to precipitation forming, and a further condition that needs to be met to sustain the process:

1 cooling of the atmosphere
2 condensation onto nuclei
3 growth of the cloud droplets/ice crystals
4 a supply of moisture to continue the process.

Atmospheric cooling

Cooling of the atmosphere may take place through several different mechanisms occurring independently or simultaneously. The most common form of cooling is from the uplift of air through the atmosphere. As described earlier, when air rises upwards into an area of lower pressure, it expands and cools as a consequence. The cooler temperature leads to less water vapour being retained by the air and conditions becoming favourable for **condensation**. The elevation at which this happens in the atmosphere is known as the *condensation level*. Think back to when you have seen the

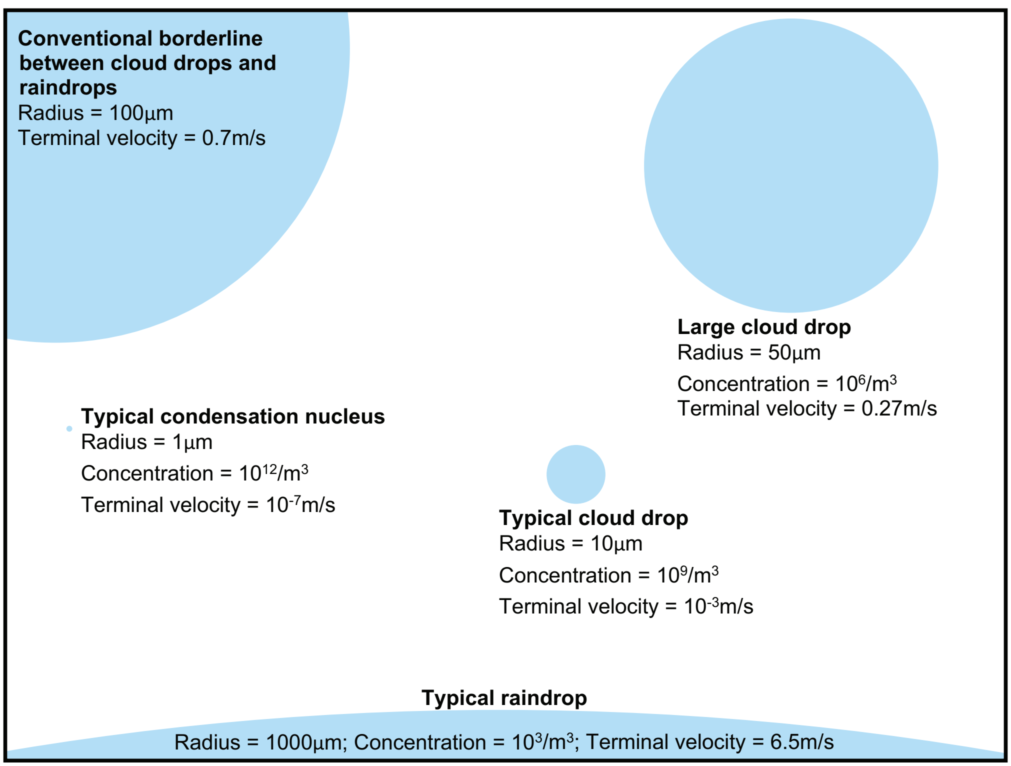

Figure 2.2 Comparative sizes, concentrations and terminal velocities of cloud droplets and raindrops.
Source: Redrawn from Sumner (1988)

sky filled with fluffy white clouds where their bases end abruptly and in what seems a straight line in the sky – this is the condensation level. The actual uplift of air may be caused by heating from the earth's surface (leading to **convective precipitation**), an air mass being forced to rise over an obstruction such as a mountain range (this leads to **orographic precipitation**), or from a low pressure weather system where the air is constantly being forced upwards (this leads to **cyclonic precipitation**). Other mechanisms whereby the atmosphere cools include a warm air mass meeting a cooler air mass, and the warm air meeting a cooler object such as the sea or land (**frontal precipitation**).

Condensation nuclei

Condensation nuclei are minute particles floating in the atmosphere (*aerosols*) which provide a surface for the water vapour to condense into liquid water upon. They are commonly less than a micron (1 micron = one-millionth of a metre or one-thousandth of a millimetre, 1 μm) in diameter (Figure 2.2). There are many different substances that make condensation nuclei, including small dust particles, sea salts and smoke particles. If condensation nuclei are not present, this is when air can become supersaturated. More often than not though, these aerosols are present and enable condensation of water vapour to microscopic cloud droplets.

Research into generating artificial rainfall has concentrated on the provision of condensation nuclei into clouds, a technique called **cloud seeding**. During the 1950s and 1960s much effort was expended in using silver iodide particles, dropped from planes, to act as condensation nuclei. However, more recent work has suggested that other salts such as potassium chloride are better nuclei. There is much controversy over the value of cloud seeding. Some studies support its effectiveness (e.g. Gagin and Neumann 1981; Ben-Zvi 1988); other authors query the results (e.g. Rangno and Hobbs 1995), while others suggest that it only works in certain atmospheric conditions and with certain cloud types (e.g. Changnon et al. 1995). More recent work in South Africa has concentrated on using hygroscopic flares to release chloride salts into the base of convective storms, with some success (Mather et al. 1997). Interestingly, this approach was first noticed through the discovery of extra heavy rainfall occurring over a paper mill in South Africa that was emitting potassium chloride from its chimney stack (Mather 1991). Countries in the Middle East are funding significant research programs in cloud seeding, as is China where it is being used to generate rain to combat urban smog.

Cloud droplet and ice particle growth

Cloud droplets or ice crystals formed around condensation nuclei are normally too small to fall directly to the ground; that is, the forces from the upward draught within a cloud are greater than the gravitational forces pulling the microscopic droplet downwards. In order to overcome the upward draughts it is necessary for the droplets to grow from an initial size of 1 micron to around 3,000 microns (3 mm). The vapour pressure difference between a cloud droplet and the surrounding air will cause it to grow through condensation, albeit rather slowly. When there are ice particles, rather than cloud droplets, the vapour pressure difference with the surrounding air becomes greater and the water vapour desublimates onto the ice particles. Desublimation occurs more quickly than condensation onto a cloud droplet, but is still a slow process. Droplets are also subject to evaporation and may undergo several evaporation-condensation cycles, or in the case of ice particles, evaporation-desublimation cycles. The main mechanism by which drops grow within a cloud is through *collision* and *coalescence*: two drops collide and join together (coalesce) to form a larger droplet that may then collide with many more before falling towards the surface as rainfall or another form of precipitation. Larger droplets are more efficient in incorporating smaller droplets because they fall through the cloud more quickly, sweeping through a larger volume of the cloud; the deeper the cloud, the greater the size of drop that is formed (Sitch and Drake 2014). When ice particles grow by collision, this is called *aggregation*, but is a very slow process.

At this point it is useful to distinguish between three types of clouds: warm, cold, and mixed clouds. Cold clouds occur at high altitudes where temperatures are less than −40 °C, and are comprised exclusively of ice particles. In contrast, warm clouds occur at lower altitudes where the temperature is greater than 0 °C and are made up of only cloud droplets (i.e. liquid water). As the name suggests, mixed clouds are somewhere between the cold and warm and have a mixture of ice and cloud particles. In mid-latitude regions of the world, mixed clouds are the most common type.

In warm clouds, the predominant mechanism for droplet growth is by collision and coalescence (see lower right of Figure 2.3). When droplets are large enough to overcome the updraught (or updraft), they will fall as rain. Between temperatures of 0 °C and as low as −40 °C, very tiny cloud droplets may not freeze, but remain as water in a supercooled state. For this to happen, no impurities should be present in the water, and neither must there be any pre-existing ice crystals. Where condensation nuclei are not present, significant supercooling in clouds can be quite common. Mixed clouds may therefore comprise both ice particles and supercooled cloud droplets at temperatures between 0 °C and −40 °C (Figure 2.3).

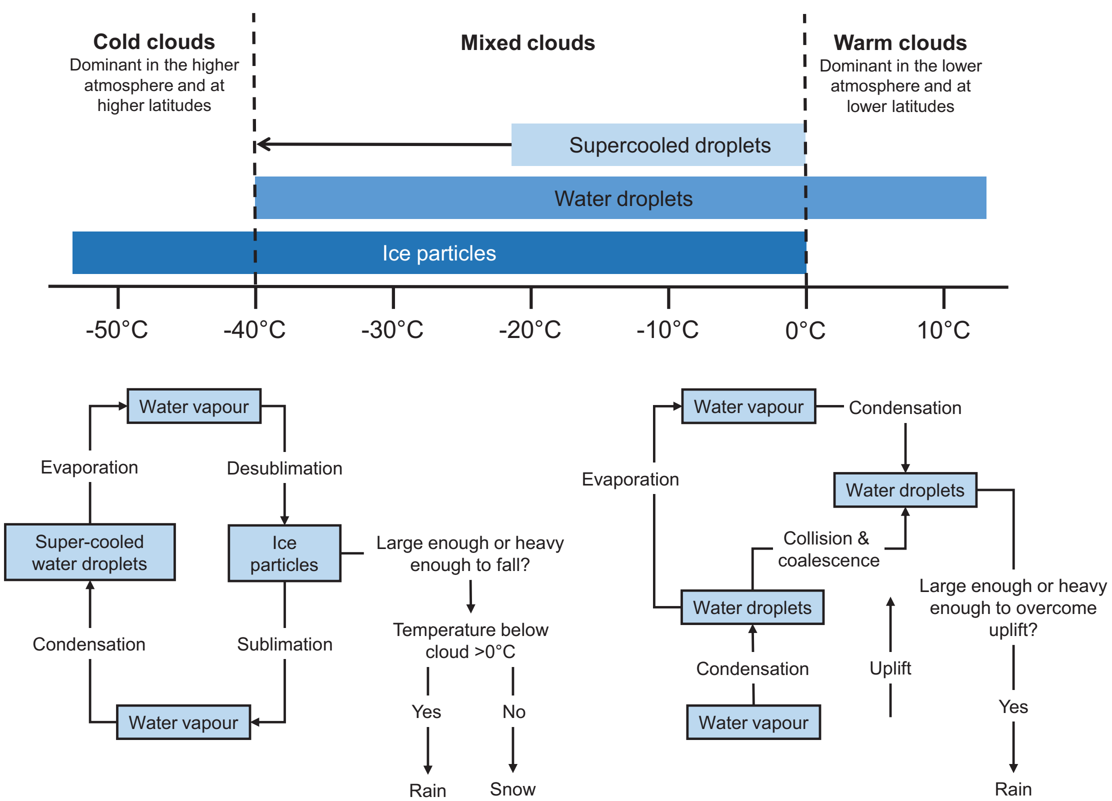

Figure 2.3 Precipitation forming processes.

Source: Adapted from Shuttleworth (2012) and Hendricks (2010)

Another important mechanism by which droplets grow sufficiently large to fall as precipitation is known as the *Bergeron-Findeisen* process, sometimes known as the cold cloud process as it can only occur when cloud temperatures are significantly below freezing. Referring back to Figure 2.1, at temperatures below freezing, there is a difference between the saturation vapour pressure curves of water and ice. This means that the air can be saturated for ice, but not for liquid water; the consequence of this is that the ice particles grow at the expense of the liquid water droplets. This happens by vapour diffusion where vapour diffuses from a higher pressure (cloud droplets) to a lower pressure (ice particles). In other words, ice particles are very efficient in attracting water vapour, which desublimates (deposits directly without liquid water) to grow the ice particles. The consequence is that the cloud becomes drier and what were cloud droplets, evaporate. This in turn provides more water vapour for desublimation, as a positive feedback process (Hendriks 2010). As the ice particles grow heavier they fall to lower in the cloud where temperatures are around 0 °C, where they clump together forming snowflakes. This process of *aggregation* is thought to be more effective where ice particles are covered by a thin film of supercooled water, making them 'sticky' as water freezes instantaneously on collision (Shuttleworth 2012). This process appears to be most effective in the temperature range −4 to 0 °C, and the largest snowflakes are produced in the warmest area of the cloud. If temperature is warmer than 0 °C below the cloud, they will melt as they fall, falling as rain.

Because clouds are usually a mixture of water vapour, liquid cloud droplets and ice crystals, the Bergeron process may be a significant factor in making cloud droplets large enough to become rain drops (or ice/snow crystals) that overcome gravity and fall out of the clouds.

The mechanisms of droplet formation within a cloud are not completely understood. The relative proportion of condensation-formed, collision-formed, and Bergeron-process-formed droplets depends very much on the individual cloud circumstances and can vary considerably. As a droplet is moved around a cloud it may freeze and thaw several times, leading to different types of precipitation (see Table 2.1).

Hail

Hail stones are typically around 5 mm, spherical, sometimes irregular in shape and comprise of ice. Larger hail stones (sometimes the size of golf balls) occur occasionally, and although very rare, even larger (>15 cm in diameter) have been recorded. Hail stones start as ice particles in the highest part of very large storm clouds (*cumulonimbus* clouds), the lower base of which are filled with supercooled water droplets. Up and down draughts in these huge clouds can be significant and when an ice particle is thrust downwards towards the base of the cloud and collides with supercooled water droplets, they freeze immediately, adding more ice. A cross-section through a typical hailstone shows alternating layers of clear and translucent ice. Where ice is opaque, freezing happened instantaneously, trapping air. Where it is clear, freezing happened more slowly (Sumner 1988). Slower freezing occurs at the tops of cumulonimbus clouds. Because of the violent up and down draughts in these very large cumulonimbus clouds, hail stones can circulate several times, before their increasing mass overcomes the up draught and they fall as hail stones. If conditions are warmer below the cloud base they may melt and fall as rain. Hail can cause serious damage to infrastructure and agriculture, and in the United States hail storms cause losses of the order of $1 billion per year (see National Storm Damage Centre https://stormdamagecenter.org/hail-storm.php).

Dew and frost

The same process of condensation occurs in **dewfall**, only in this case the water vapour condenses into liquid water (dew) after coming into contact with a cold surface. In humid-temperate countries, dew is a common occurrence in autumn/fall when the air at night is still warm but vegetation and

Table 2.1 Types of precipitation

Class	*Definition*
Drizzle	A subset of fine rain with droplets between 0.1 and 0.5 mm, but close together
Rain	Liquid water droplets with diameter between 0.5 and 0.7 mm, but smaller if widely scattered
Freezing rain or drizzle	Rain or drizzle, the drops of which freeze on impact with a solid surface. Also called sleet in the USA
Sleet	Partly melted snowflakes, or rain and snow falling together (UK). Fairly transparent grains or pellets of ice (USA)
Ice crystals, ice prisms, snow and snowflakes	Snow can fall as single branched hexagonal or star-like ice crystals, or in the case of ice prisms, as unbranched ice crystals in the form of hexagonal needles, columns or plates. The nature of the crystal depends on the temperature at which it forms and the corresponding amount of water vapour. More often snow falls as agglomerated snowflakes.
Snow grains	Very small, white, opaque grains of ice, flat or elongated, with diameter generally <1 mm. Also called granular snow
Snow pellets	White, opaque grains of spherical or conical ice (2–5 mm). Also called granular snow, or graupel
Ice pellets	Transparent or translucent pellets of ice, spherical or irregular with diameter <5 mm
Hail	Balls or pieces of ice usually between 5 and 125 mm in diameter, commonly showing alternating concentric layers of clear and opaque ice in cross-section

Source: Adapted from Shuttleworth (2012) and Sumner (1988)

other surfaces have cooled to the point where water vapour coming into contact with them condenses onto the leaves and forms dew. Dew is not normally a major part of the hydrological cycle but is another form of precipitation. Frost occurs in a similar way; either through later freezing of dewfall or when the surface on which it forms is less than 0 °C, so deposition or desublimation occurs, rather than condensation. Deposition of crystals can form intricate forms, called *hoar* frost.

Mist and fog

Mist and fog occur under similar conditions to dewfall, and dew is often their precursor (Sumner 1988). Cooling air and condensation of dew causes a shallow layer of saturated air near the surface. Condensation nuclei in this layer of air enable mist to form and progressively deepen and thicken, ultimately forming a fog where visibility is impaired (usually <200 m on land. Several types of fog can be recognised (Table 2.2), and fog in arid environments can be an important source of moisture for biological processes.

PRECIPITATION DISTRIBUTION

The amount of precipitation falling over a location varies both spatially and temporally (with time). The different influences on the precipitation can be divided into static and dynamic influences. Static influences are those such as altitude, aspect and slope; they do not vary between storm events. Dynamic influences are those that do change and are by and large caused by variations in the weather. At the global scale the influences on precipitation distribution are mainly dynamic, being caused by differing weather patterns, but there are static factors such as topography that can also cause major variations through a **rain shadow effect**

Table 2.2 Types of fog

Type of fog	*Description*
Radiation fog	During the day the earth absorbs heat but radiates outwards at night, cooling the land and lower air layers. If sufficient cooling occurs fog can develop. Ideal conditions include clear skies on long nights and slight winds to distribute the cold air over wider areas. This is the most common type of fog, usually occurring in valleys and typically around sunrise.
Advection fog	When warmer, moist air blows over a colder surface, the lower layers can cool sufficiently to condense. The fog often pictured around the Golden Gate Bridge in San Francisco forms in this way.
Coastal fog or sea fog	A type of advection fog where a warmer, moist flow of air passes over the colder sea, cooling the lower layers until condensation occurs. Often found along the Californian coast, and the west coast of the Namibian desert where it is an important source of moisture.
Hill or upslope fog	Moist warmer air flows up the windward slope, where it cools and condenses.
Freezing fog	Comprised of supercooled water droplets at temperatures well below freezing. *Rime* occurs where supercooled droplets instantaneously freeze on contact with solid objects (e.g. telephone wires, trees)
Smog	Air pollution creates an abundance of condensation nuclei, so a thick fog can form under ideal conditions. Such conditions occurred regularly in the nineteenth and early twentieth century in London.

(see Case Study, pp. 29–30). At the continental scale, large differences in rainfall can be attributed to a mixture of static and dynamic factors. In Figure 2.4 the rainfall distribution across the USA shows marked variations. Although mountainous areas have a higher rainfall, and also act as a block to rainfall reaching the drier centre of the country, they do not provide the only explanation for the variations evident in Figure 2.4. The higher rainfall in the north-west states (Oregon and Washington) is linked to wetter cyclonic weather systems from the northern Pacific that do not reach down to southern California. Higher rainfall in Florida and other southern states is linked to the warm waters of the Caribbean Sea. These are examples of dynamic influences as they vary between rainfall events.

At smaller scales, the static factors are often more dominant, although it is not uncommon for quite large variations in rainfall across a small area caused by individual storm clouds to exist. As an example: on 3 July 2000 an intense rainfall event caused flooding in the village of Epping Green, Essex, UK. Approximately 10 mm of rain fell within 1 hour, although there was no recorded rainfall in the village of Theydon Bois approximately 10 km to the south. This large spatial difference in rainfall was caused by the scale of the weather system causing the storm – in this case a convective thunderstorm. Often these types of variation lessen in importance over a longer timescale so that the annual rainfall in Epping Green and Theydon Bois is very similar, whereas the daily rainfall may differ considerably. For the hydrologist, who is interested in rainfall at the small scale, the only way to try and characterise these dynamic variations is through having as many **rain gauges** as possible within a study area.

Static influences on precipitation distribution

It is easier for the hydrologist to account for static variables such as those discussed below.

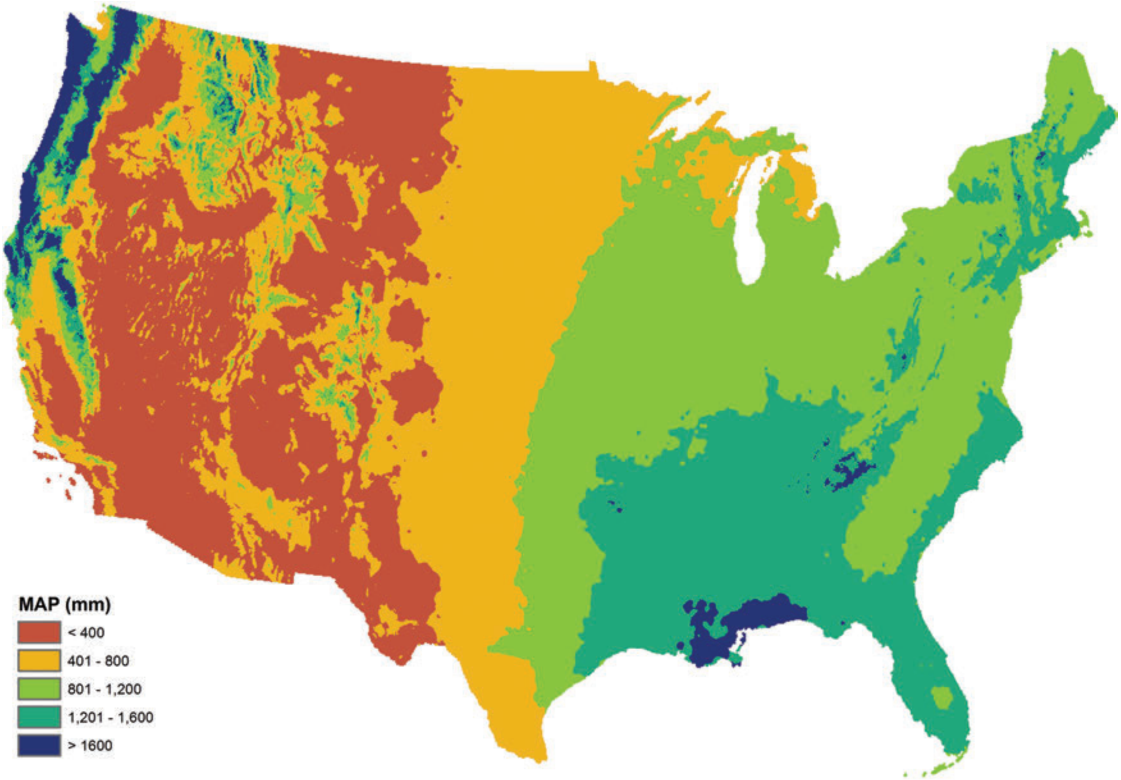

Figure 2.4 Mean annual precipitation across the USA (1981–2010).

Source: Mapped using data from PRISM Climate Group, Oregon State University, http://prism.oregonstate.edu

Altitude

It has already been explained that temperature is a critical factor in controlling the amount of water vapour that can be held by air. The cooler the air is, the less water vapour can be held. As temperature decreases with altitude it is reasonable to assume that as an air parcel gains altitude it is more likely to release the water vapour and cause higher rainfall. This is exactly what does happen and there is a strong correlation between altitude and rainfall: so-called *orographic precipitation*.

Aspect

The influence of aspect is less important than altitude but it may still play an important part in the distribution of precipitation throughout a catchment. In the humid mid-latitudes (35° to 65° north or south of the equator) the predominant source of rainfall is through cyclonic weather systems arriving from the west. Slopes within a catchment that face eastwards will naturally be more sheltered from the rain than those facing westwards. The same principle applies everywhere: slopes with aspects facing away from the predominant weather patterns will receive less rainfall than their opposites.

Slope

The influence of slope is only relevant at a very small scale. Unfortunately the measurement of rainfall occurs at a very small scale (i.e. a rain gauge). The difference between a level rain gauge on a hillslope, compared to one parallel to the slope, may be significant. It is possible to calculate this difference if it is assumed that rain falls vertically – but of course rain does not always fall vertically. Consequently the effect of slope on rainfall measurements is normally ignored.

Rain shadow effect

Where there is a large and high land mass it is common to find the rainfall considerably higher on one side than the other. This is through a combination of altitude, slope, aspect and dynamic weather direction influences and can occur at many different scales (see Case Study below).

Case study

THE RAIN SHADOW EFFECT

The predominant weather pattern for the South Island of New Zealand is a series of rain-bearing depressions sweeping up from the Southern Ocean, interrupted by drier blocking anticyclones. The Southern Alps form a major barrier to the fast-moving depressions and have a huge influence on the rainfall distribution within the South Island. Formed as part of tectonic uplift along the Pacific/Indian plate boundary, the Southern Alps stretch the full length of the South Island (approximately 700 km) and at their highest point are over 3,000 m above mean sea level.

The predominant weather pattern has a westerly airflow, bringing moist air from the Tasman Sea onto the South Island. As this air is forced up over the Southern Alps it cools down and releases much of its moisture as rain and snow. As the air descends on the eastern side of the mountains it warms up and becomes a föhn wind, referred to locally as a 'nor-wester'. The annual rainfall patterns for selected stations in the South Island are shown in Figure 2.5. The rain shadow effect can be clearly seen with the west coast rainfall being at least four times that of the east. Table 2.3 also illustrates the point, with the number of rain days at different sites in a cross section across the South Island. Although not shown on the transect in Figure 2.5, recordings of rainfall further north in the Southern Alps (Cropp River inland from Hokitika) are as high as 6 m a year.

This pattern of rain shadow is seen at many different locations around the globe. It does not

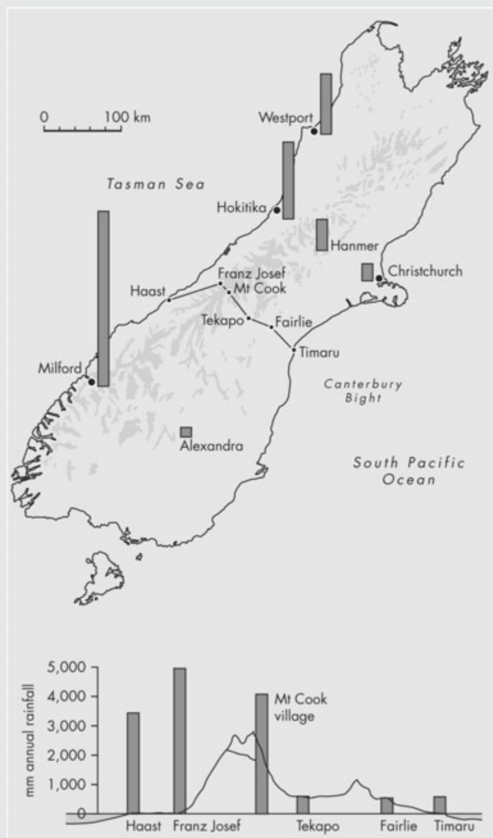

Figure 2.5 Rainfall distribution across the Southern Alps of New Zealand (South Island). Shaded areas on the map are greater than 1,500 m in elevation. A clear rain shadow effect can be seen between the much wetter west coast and the drier east.

require as large a barrier as the Southern Alps – anywhere with a significant topographical barrier is likely to cause some form of rain shadow. Hayward and Clarke (1996) present data showing a strong rain shadow across the Freetown Peninsula in Sierra Leone. They analysed mean monthly rainfall in 31 gauges within a 20 × 50 km area, and found that the rain shadow effect was most marked during the monsoon months of June to October. The gauges in locations facing the ocean (south-west aspect) caught considerably more rainfall during the monsoon than those whose aspect was towards the north-east and behind a small range of hills.

Table 2.3 Average annual rainfall and rain days for a cross section across South Island

Weather station	*Height above mean sea level*	*Annual rainfall (mm)*	*Rain days*
Haast	30	5,840	175
Mt Cook village	770	670	120
Tekapo	762	604	77
Timaru	25	541	75

Note: More details on weather differentials across the South Island of New Zealand are in Sinclair et al. (1996)
Source: Data from New Zealand Met. Service and other miscellaneous sources

MEASUREMENT

For hydrological analysis it is important to know how much precipitation has fallen and when this occurred. The usual expression of precipitation is as a vertical depth of liquid water. Rainfall is measured by millimetres or inches depth, rather than by volume such as litres or cubic metres. The measurement is the depth of water that would accumulate on the surface if all the rain remained where it had fallen (Shaw 1994). Snowfall may also be expressed as a depth, although for hydrological purposes it is most usefully described in water equivalent depth (i.e. the depth of water that would be present if the snow melted). This is a recognition that snow takes up a greater volume (as much as 90 per cent more) for the same amount of liquid water.

There is a strong argument that can be made to say that there is no such thing as precipitation measurement at the catchment scale as it varies so tremendously over a small area. The logical end-point to this argument is that all measurement techniques are in fact precipitation estimation techniques. For the sake of clarity in this text, precipitation measurement techniques refer to the methods used to quantify the volume of water present, as opposed to estimation techniques where another variable is used as a surrogate for the water volume.

Rainfall measurement

The earliest written reference to measuring rainfall dates from the fourth century BC in India (Strangeways 2007), but systematic measurement of rainfall was in place in Korea by the mid-1400s (Figure 2.6a). The instrument for measuring rainfall is called a *rain gauge*. A rain gauge measures the volume of water that falls onto a horizontal surface delineated by the rain gauge rim (see Figure 2.6b). The collected volume is converted into a rainfall depth through division by the rain gauge surface area. The modern rain gauge has its origins in the designs of Sir Christopher Wren and Robert Hooke (Strangeways 2007). The design of a rain gauge is not as simple as it may seem at first glance as there are many errors and inaccuracies that need to be minimised or eliminated. The World Meteorological Organisation (WMO) suggests that errors are typically between 3 per cent and 30 per cent or more, depending on the gauge and weather and site conditions (WMO 2008), and errors are generally accepted to be of the order of 5–10 per cent, but higher for snow (Shuttleworth

Figure 2.6 (a) A fourteenth-century rain gauge from Korea. Photo: Tim Davie. (b) A rain gauge sitting above the surface to avoid splash.

2012). When up-scaled to the catchment area, this represents significant uncertainty in the major input of the water balance equation.

There is a considerable scientific literature studying the accuracy and errors involved in measuring rainfall. It needs to be borne in mind that a rain gauge represents a very small point measurement (or sample) from a much larger area that is covered by the rainfall. Any errors in measurement will be amplified hugely because the rain gauge collection area represents such a small sample size. Because of this amplification it is extremely important that the design of a rain gauge negates any errors and inaccuracies.

The four main sources of error in measuring rainfall that need consideration in designing a method for the accurate measurement of rainfall are:

1 losses due to evaporation
2 losses due to wetting of the gauge
3 over-measurement due to splash from the surrounding area
4 under-measurement due to turbulence around the gauge.

Evaporation losses

A rain gauge can be any collector of rainfall with a known collection area; however, it is important that any rainfall that does collect is not lost again through evaporation. In order to eliminate, or at least lessen this loss, rain gauges are funnel shaped. In this way the rainfall is collected over a reasonably large area and then any water collected is passed through a narrow aperture to a collection tank underneath. Because the collection tank has a narrow top (i.e. the funnel mouth) there is very little interchange of air with the atmosphere above the gauge. As will be explained in Chapter 3, one of the necessary requirements for evaporation is the turbulent mixing of saturated air with drier air above. By restricting this turbulent transfer there is little evaporation that can take place. In addition to this, the water awaiting measurement is kept out of direct sunlight so that it will not be warmed; hence there is a low evaporation loss. Evaporation losses are estimated to be of the order of 0–4 per cent (WMO 2008) and would depend on meteorological factors such as the type of precipitation, the prevailing capacity of the air to hold water and the wind speed at the rim of the gauge over the period between rain collecting and a measurement being recorded. Instrumental factors would include for example, the funnel orifice area and colour of the gauge (WMO 2008).

Wetting loss

As the water trickles down the funnel it is inevitable that some water will stay on the surface of the funnel and can be lost to evaporation or not measured in the collection tank. This is often referred to as a *wetting loss*. These losses will not be large but may be significant, particularly if the rain is falling as a series of small events on a warm day. In order to lessen this loss it is necessary to have steep sides on the funnel and to have a non-stick surface. The standard UK Meteorological Office rain gauge is made of copper to create a non-stick surface, although many modern rain gauges are made of non-adhesive plastics. Wetting loss is thought to be in the range of 2–10 per cent (WMO 2008), and meteorological factors include the type and frequency of precipitation, relative to the drying time of the gauge and frequency of emptying. Gauge factors include the material and age of the gauge and collector (WMO 2008).

Rain splash

The perfect rain gauge should measure the amount of rainfall that would have fallen on a surface if the gauge was not there. This suggests that the ideal situation for a rain gauge is flush with the surface. A difficulty arises, however, as a surface-level gauge is likely to over-measure the catch due to rain landing adjacent to the gauge and splashing into it. If there was an equal amount of splash going out of the gauge then the problem might not be so severe, but the sloping sides of the funnel (to reduce evaporative losses) mean that there will be very little splash-out. In extreme situations it is even possible that the rain gauge could be flooded by water flowing over the surface or covered by snow. To overcome the splash, flooding and snow coverage problem the rain gauge can be raised up above the ground (Figure 2.6) or placed in the middle of a non-splash grid (see Figure 2.7). Splash-in error is around 1–2 per cent (WMO 2008), primarily a function of rainfall intensity and wind speed. Gauge factors include the depth and shape of the gauge, and type of installation (WMO 2008).

Turbulence around a raised gauge

If a rain gauge is raised up above the ground (to reduce splash) another problem is created due to air turbulence around the gauge. The rain gauge presents an obstacle to the wind and the consequent aerodynamic interference leads to a reduced catch (see Figure 2.8). The amount of loss is dependent

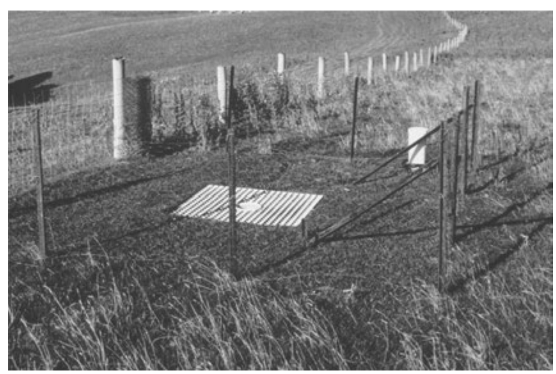

Figure 2.7 Surface rain gauge with non-splash surround.

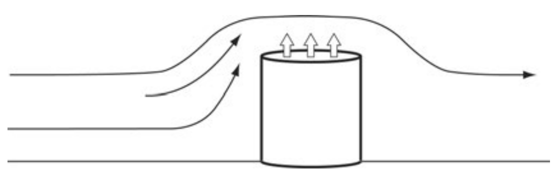

Figure 2.8 The effect of wind turbulence on a raised rain gauge. An area of reduced pressure (and uplift) develops above the gauge in a similar manner to an aircraft wing. This reduces the rain gauge catch.

on both the wind speed and the raindrop diameter (Nešpor and Sevruk 1999). At wind speeds of 20 km/hr (Beaufort scale 2) the loss could be up to 20 per cent, and in severe winds of 90 km/hr (Beaufort scale 8) up to 40 per cent (Bruce and Clark 1980; Rodda and Smith 1986). A generalised loss of 2.2 per cent for every 1 m/s increase in wind speed has been found (Larson and Peck 1974). The higher a gauge is from the surface, the greater the loss of accuracy. This creates a major problem for gauges in areas that receive large snowfalls as they need to be raised to avoid surface coverage.

One method of addressing these turbulence difficulties is through the fitting of a shield to the rain gauge (see Figure 2.9). A rain gauge shield can take many forms (two examples are a Nipher shield and an Alter shield) but is often a series of batons surrounding the gauge at its top height. The shield acts as a calming measure for wind around the gauge and has been shown to greatly improve rain gauge accuracy.

The optimum rain gauge design

There is no perfect rain gauge. The design of the best gauge for a site will be influenced by the individual conditions at the site (e.g. prevalence of snowfall, exposure, etc.). A rain gauge with a non-splash surround, such as in Figure 2.7, can give very accurate measurement but it is prone to coverage by heavy snowfall so cannot always be used. The non-splash surround allows adjacent rainfall to pass through (negating splash) but acts as an extended soil surface for the wind, thereby eliminating the turbulence problem from raised gauges. This may be the closest that it is possible to get to measuring the amount of rainfall that would have fallen on a surface if the rain gauge were not there.

The standard UK Meteorological Office rain gauge has been adopted around the world (although not everywhere) as a compromise between the factors influencing rain gauge accuracy. It is a brass-rimmed rain gauge of 5 inches (127 mm) diameter standing 1 foot (305 mm) above the ground. The lack of height above ground level is a reflection of the low incidence of snowfall in the UK; in countries such as Russia and Canada, where winter snowfall is the norm, gauges may be raised as high as 2 m above the surface. There is general recognition that the UK standard rain gauge is not the best design for hydrology, but it does represent a reasonable compromise. There is a strong argument to be made against changing its design. Any change in the measurement instrument would make an analysis of past rainfall patterns difficult due to the differing accuracy.

Figure 2.9 Baffles surrounding a rain gauge to lessen the impact of wind turbulence. The gauge is above ground because of snow cover during the winter.

Siting of a rain gauge

Once the best measurement device has been chosen for a location there is still a considerable measurement error that can occur through incorrect siting. The major problem of rain gauge siting in hydrology is that the scientist is trying to measure the rainfall at a location that is representative of a far greater area. It is extremely important that the measurement location is an appropriate surrogate for the larger area. If the area of interest is a forested catchment then it is reasonable to place your rain gauge beneath the forest canopy; likewise, within an urban environment it is reasonable to expect interference from buildings because this is what is happening over the larger area. What is extremely

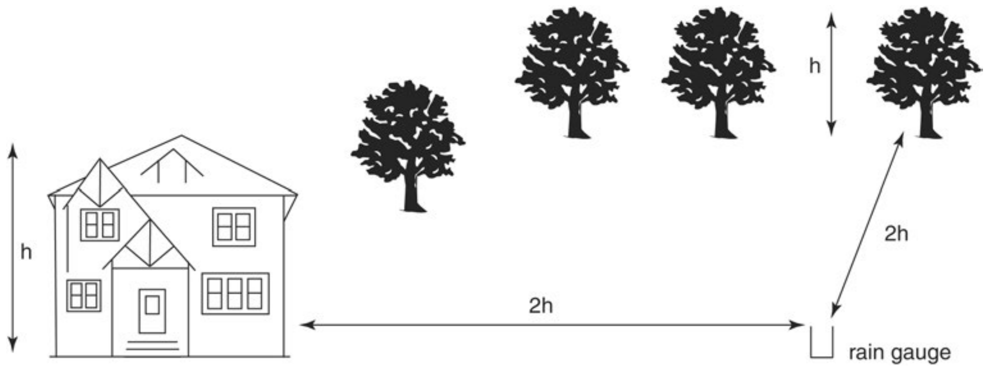

Figure 2.10 Siting of a rain gauge away from obstructions.

important is that there are enough rain gauges to try to quantify the spatial and temporal variations.

The rule-of-thumb method for siting a rain gauge is that the angle when drawn from the top of the rain gauge to the top of the obstacle is less than 30° (see Figure 2.10). This can be approximated as at least twice the height of the obstacle away from the gauge. Care needs to be taken to allow for the future growth of trees so that at all times during the rainfall record the distance apart is at least twice the height of an obstacle.

Gauges for the continuous measurement of rainfall

The standard UK Meteorological Office rain gauge collects water beneath its funnel and this volume is read once a day. Often in hydrology the data needs to be measured at a finer timescale than this, particularly in the case of individual storms which often last much less than a day. The most common modern method for measuring continuous rainfall uses a tipping-bucket rain gauge. These are very simple devices that can be installed relatively cheaply, although they do require a data-logging device nearby. The principle behind the tipping-bucket rain gauge is that as the rain falls it fills up a small 'bucket' that is attached to another 'bucket' on a balanced cross arm (see Figure 2.11). The 'buckets' are very small plastic containers at the end of each cross arm. When the bucket is full it tips the balance so that the full bucket is lowered down and empties out. At the time of tipping, a magnet attached to the balance arm closes a small reed switch which sends an electrical signal to a data-logging device. This then records the exact time of the tipped bucket. If the rain continues to fall it fills the bucket on the other end of the cross arm until it too tips the balance arm, sending another electrical impulse to the data logger. In this way a near continuous measurement of rainfall with time can be obtained.

Figure 2.11 The inside of a tipping-bucket rain gauge. The 'buckets' are the small white, triangular reservoirs. These are balanced and when full they tip over, bringing the black arm past the other stationary arm. In doing so a small electrical current is passed to a data logger.

It is important that the correct size of tipping bucket is used for the prevailing conditions. If the buckets are too small then a very heavy rainfall event will cause them to fill too quickly and water be lost through overspill while the mechanism tips. If the buckets are too large then a small rainfall event may not cause the cross arm to tip and the subsequent rainfall event will appear larger than it actually was. The tipping-bucket rain gauge shown in Figure 2.11 has an equivalent depth of 0.2 mm of rain which works well for field studies in south-east England, but would be inadequate for a tropical downpour.

Snowfall measurement

The measurement of snowfall has similar problems to those presented by rainfall, but they are often more extreme. There are two methods used for measuring snowfall: using a gauge like a rain gauge; or measuring the depth that is present on the ground. Both of these methods have very large errors associated with them, predominantly caused by the way that snow falls through the atmosphere and is deposited on the gauge or ground. Most, although not all, snowflakes are more easily transported by the wind than raindrops are. When the snow reaches the ground it is easily blown around in a secondary manner (drifting). This can be contrasted to liquid water where, upon reaching the ground, it is either absorbed by the soil or moves across the surface. Rainfall is very rarely picked up by the wind again and redistributed in the manner that drifting snow is. For the snow gauge this presents problems that are analogous to rain splash. For the depth gauge the problem is due to uneven distribution of the snow surface: it is likely to be deeper in certain situations.

Rain gauge modification to include snowfall

One modification that needs to be made to a standard rain gauge in order to collect snowfall is a heated rim so that any snow falling on the gauge melts to be collected as liquid water. Failure to have a heated rim may mean that the snow builds up on the gauge surface until it overflows. Providing a heated rim is no simple logistical exercise as it necessitates a power source (difficult in remote areas) and the removal of collected water well away from the heat source to minimise evaporation losses.

A second modification is to raise the gauge well above ground level so that as snow builds up the gauge is still above this surface. Unfortunately the raising of the gauge leads to an increase in the turbulence errors described for rain gauges. For this reason it is normal to have wind deflectors or shields surrounding the gauge.

Snow depth

The simplest method of measuring snow depth is the use of a core sampler. This takes a core of snow, recording its depth at the same time, that can then be melted to derive the water equivalent depth. It is this that is of importance to a hydrologist. The major difficulties of a core sample are that it is a non-continuous reading (similar to daily rainfall measurement), and the position of coring may be critical (because of snow drifting).

A second method of measuring snow depth is to use a **snow pillow**. This is a method for measuring snow accumulation, a form of water storage, hence it is described in Chapter 4 (p. 81).

Summary of the challenges in point measurement

As is evident from the preceding discussion, the estimation of precipitation at a point is challenging. Figure 2.12 summarises a range of possible errors influencing catch recorded at a site. In the next section we will move on to how one might generalise point rainfall over a wider area and this process has even greater shortfalls, further compounding the total error in estimating precipitation.

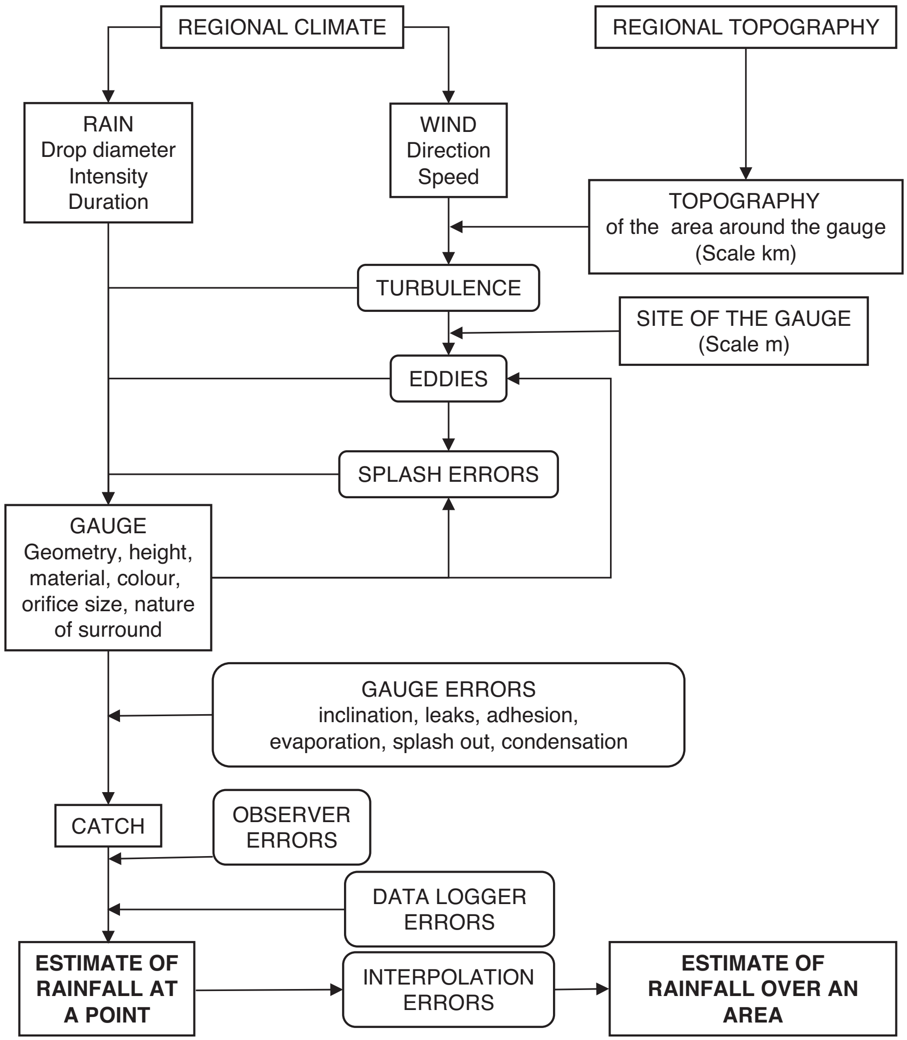

Figure 2.12 Potential sources of error in measurement of rainfall at a point and over an area.

Source: Adapted from Rodda (1967)

MOVING FROM POINT MEASUREMENT TO SPATIALLY DISTRIBUTED ESTIMATION

The measurement techniques described here have all concentrated on measuring rainfall at a precise location (or at least over an extremely small area). In reality, the hydrologist needs to know how much precipitation has fallen over a far larger area, usually a catchment. To move from point measurements to a spatially distributed estimation it is necessary to employ some form of spatial averaging. The spatial averaging must attempt to account for an uneven spread of rain gauges in the catchment and the various factors that we know influence rainfall distribution (e.g. altitude, aspect and slope). A simple arithmetic mean will only work where a catchment is sampled by uniformly spaced rain gauges and where there is no diversity in topography. If these conditions were ever truly met then it is unlikely that there would be more than one rain gauge sampling the area. Hence it is very rare to use a simple averaging technique.

There are different statistical techniques that address the spatial distribution issues, and with the growth in use of **Geographic Information Systems (GIS)** it is often a relatively trivial matter to do the calculation. As with any computational task it is important to have a good knowledge of how the technique works so that any shortcomings are fully understood. Three techniques are described here: **Thiessen's polygons**, the **hypsometric method** and the **isohyetal method**. These methods are explored further in a Case Study on p. 40.

Thiessen's polygons

Thiessen was an American engineer working around the start of the twentieth century who devised a simple method of overcoming an uneven distribution of rain gauges within a catchment (very much the norm). Essentially Thiessen's polygons (Thiessen 1911) attach a representative area to each rain gauge. The size of the representative area (a polygon) is based on how close each gauge is to the others surrounding it, but all points within a polygon are closer to its rain gauge than any of the other rain gauges.

Polygons are drawn by connecting the nearest rain gauges to each other by lightly drawn lines. The perpendicular bisector of each connecting line is then found, and these are extended to where they intersect with other perpendicular bisectors. The boundaries of the polygons are therefore equidistant from each gauge (see Figure 2.13). Once the polygons have been drawn, the area of each polygon surrounding a rain gauge is found. The spatially averaged rainfall (R) is calculated using Formula 2.1:

$$R = \sum_{i=1}^{n} r_i \left(\frac{a_i}{A} \right) \qquad (2.1)$$

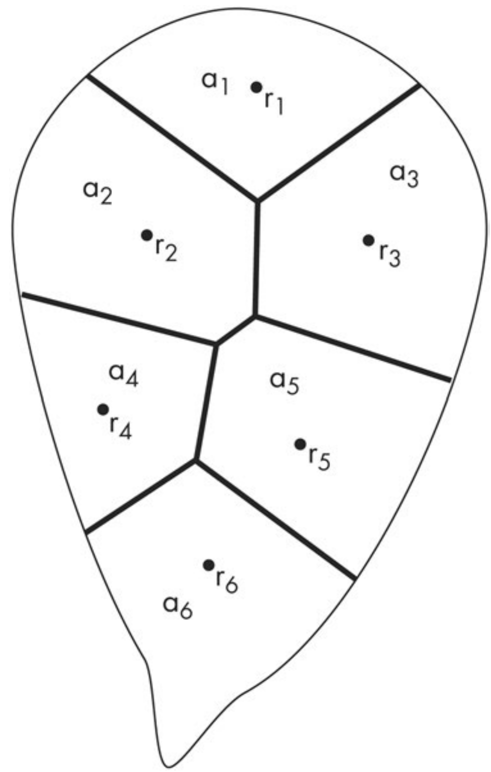

Figure 2.13 Thiessen's polygons for a series of rain gauges (r_i) within an imaginary catchment. The area of each polygon is denoted as a_i. Locations of rain gauges are indicated by bullet points.

where r_i is the rainfall at gauge i, a_i is the area of the polygon surrounding rain gauge i, and A is the total catchment area (a_i/A is therefore the proportion of the catchment occupied by each polygon, and the set of these for a catchment are known as Thiessen coefficients).

The **areal rainfall** value using Thiessen's polygons is a weighted mean, with the weighting being based upon the size of each representative area (polygon). This technique is only truly valid where the topography is uniform within each polygon so that it can be safely assumed that the rainfall distribution is uniform within the polygon. This would suggest that it can only work where the rain gauges are located initially with this technique in mind (i.e. *a priori*).

Hypsometric method

Since it is well known that rainfall is positively influenced by altitude (i.e. the higher the altitude the greater the rainfall) it is reasonable to assume that knowledge of the catchment elevation can be brought to bear on the spatially distributed rainfall estimation problem. The simplest indicator of the catchment elevation is the hypsometric (or hypsographic) curve. This is a graph showing the proportion of a catchment above or below a certain elevation. The values for the curve can be derived from maps using a planimeter or using a digital elevation model (DEM) in a GIS.

The hypsometric method of calculating spatially distributed rainfall then calculates a weighted average based on the proportion of the catchment between two elevations and the measured rainfall between those elevations (Equation 2.2).

$$R = \sum_{i=1}^{n} r_j \left(\frac{a_i}{A} \right) \tag{2.2}$$

where r_j is the average rainfall between two contour intervals, a_i is the area between those contours (derived from the hypsometric curve), and A is the total catchment area. So a_i/A is again the proportion of the catchment, but this time on the basis of the area between the contours rather than each polygon, as was the case in the previous method. The r_j value may be an average of several rain gauges where there is more than one at a certain contour interval. This is illustrated in Figure 2.14 where the shaded area (a_3) has two gauges within it. In this case the r_j value will be an average of r_4 and r_5.

Intuitively this is producing representative areas for one or more gauges based on contours and spacing, rather than just on the latter as for Thiessen's polygons. There is an inherent assumption that elevation is the only topographic parameter affecting rainfall distribution (i.e. slope and aspect are ignored). It also assumes that the relationship between altitude and rainfall is linear, which is not always the case and warrants exploration before using this technique.

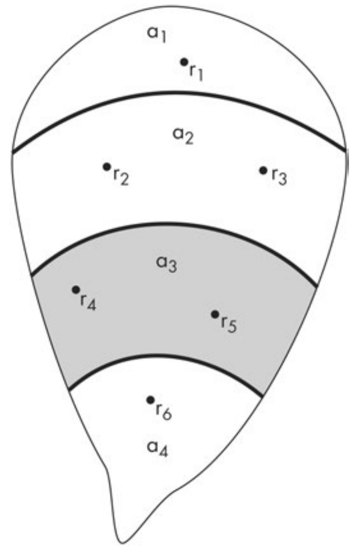

Figure 2.14 Calculation of areal rainfall using the hypsometric method. The shaded region is between two contours. In this case the rainfall is an average between the two gauges within the shaded area. Locations of rain gauges are indicated by bullet points.

Isohyetal method

Where there are a large number of gauges within a catchment the most obvious weighting to apply on a mean is based on measured rainfall distribution rather than on surrogate measures as described above. In this case a map of the catchment rainfall distribution can be drawn by interpolating between the rainfall values, creating a smoothed rainfall surface. The traditional isohyetal method involved drawing isohyets (lines of equal rainfall) on the map and calculating the area between each isohyet. The spatial average could then be calculated by Equation 2.3:

$$R = \sum_{i=1}^{n} r_j \left(\frac{a_i}{A} \right) \tag{2.3}$$

where a_i is the area between each isohyet and r_i is the average rainfall between the isohyets. This technique is analogous to Figure 2.14, except in this case the contours will be of rainfall rather than elevation.

GIS, interpolation and other spatial statistical techniques

If you undertook manual interpolation in applying the isohyetal method above, you would be undertaking linear interpolation, where the interpolated value between two known points is directly proportional to the distance between them. With the advent of GIS, interpolating and drawing of isohyets can be done relatively easily, and GIS offers a wide range of methods beyond simple linear interpolation. GIS-based methods generally subdivide the catchment into small grid cells and then assign a rainfall value for each grid cell to derive a rainfall surface by smoothing or interpolation. In both approaches measured values are used to generate a surface, but in the case of smoothing methods the surface does not necessarily pass through the points, whereas in the case of interpolation it does (Dingman 2008). There are several different ways of carrying out the interpolation, for example, the inverse distance, or nearest neighbour methods. In nearest neighbour interpolations, the assigned rainfall value for a grid square is proportional to the nearest rain gauges. In the case of the inverse distance method, the closer a rainfall station is to a point the greater the relative weighting; interpolation is therefore on a non-linear basis. Surface smoothing approaches such as the spline method fit a surface using low order polynomial functions. These approaches are also referred to as multi-quadric surface methods (Shaw et al. 2011).

While smoothing and interpolation are examples of one type of approach to developing a surface from a number of point estimates, there is an alternative category of approaches based on geostatistics. These techniques have their origin in mining applications where, for example, surfaces needed to be generated on the basis of point samples of mineral concentrations in ore. They are well-suited to any analysis where there is *spatial correlation*, which is where the difference between values at pairs of points tends to reduce with increasing proximity (i.e. the closer they are together the more likely they are to be similar). One approach is called **kriging**, which comprises a two stage process. The first step in this process is to establish how 'predictable' the values are at each sampled point from all other sampled points. This analysis results in a special type of graph known as a *semi-variogram* which represents the differences between the known value at one location and the value at a different location based on the distance (and sometimes the direction) between them. This defines the *covariance* of points, a measure of how much two random variables change together. The second stage is where the value of points at unsampled locations (where values are not known), are estimated. A weighted average of neighbouring known points is used to estimate the unsampled points, with the weights being determined on the basis of information from the semi-variogram. *Co-kriging* is a variation of this approach where other variables (e.g. elevation) can be included (Shaw et al. 2011). What is particularly useful about these methods, is that they can also provide an estimate of the uncertainty of estimation. A fuller explanation of these techniques is provided by Bailey and Gatrell (1995).

An additional piece of information that can be gained from interpolated rainfall surfaces is the likely rainfall at a particular point within the catchment. This may be more useful information than total rainfall over an area, particularly when needed for numerical simulation of hydrological processes.

The difficulty in moving from the point measurement to a spatially distributed average is a prime example of the problem of scale that besets hydrology. The scale of measurement (i.e. the rain gauge surface area) is far smaller than the catchment area that is frequently our concern. Is it feasible to simply scale up our measurement from point sources to the overall catchment? Or should there be some form of scaling factor to acknowledge the large discrepancy? There is no easy answer to these questions and they are the type of problem that research in hydrology will be investigating in the twenty-first century.

Case study

RAINFALL DISTRIBUTION IN A SMALL STUDY CATCHMENT

It is well known that large variations in rainfall occur over quite a small spatial scale. Despite this, there are not many studies that have looked at this problem in detail. One study that has investigated spatial variability in rainfall was carried out in the Plynlimon research catchments in mid-Wales (Clarke et al. 1973). In setting up a hydrological monitoring network in the Wye and Severn catchments, 38 rain gauges were installed to try and characterise the rainfall variation. The rainfall network had 18 rain gauges in the Severn catchment (total area 8.7 km^2) and 20 gauges in the Wye (10.55 km^2).

The monthly data for a period between April 1971 and March 1973 were analysed to calculate areal average rainfall using contrasting methods. The results from this can be seen in Figure 2.15. The most startling feature of Figure 2.15 is the lack of difference in calculated values and that they follow no regular pattern. At times the arithmetic mean is greater than the others while in other months it is less. When the total rainfall for the 2-year period is looked at, the Thiessen's calculation is 0.3 per cent less than the arithmetic mean, while the isohyetal method is 0.4 per cent less.

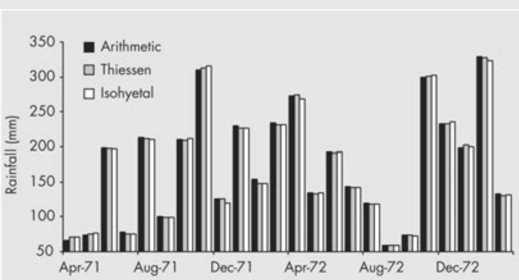

Figure 2.15 Areal mean rainfall (monthly) for the Wye catchment, calculated using three different methods.

Source: Data from Clarke et al. (1973)

When the data were analysed to see how many rain gauges would be required to characterise the rainfall distribution fully it was found that the number varied with the time period of rainfall and the season being measured. When monthly data were looked at there was more variability in winter rainfall than summer. For both winter and summer it showed that anything less than five rain gauges (for the Wye) increased the variance markedly.

A more detailed statistical analysis of hourly mean rainfall showed a far greater number of gauges were required. Four gauges would give an accuracy in areal estimate of around 50 per cent, while a 90 per cent accuracy would require 100 gauges (Clarke et al. 1973: 62).

The conclusions that can be drawn from the study of Clarke et al. (1973) are of great concern to hydrology. It would appear that even for a small catchment a large number of rain gauges are required to try and estimate rainfall values properly. This confirms the statement made at the start of this chapter: although rainfall is relatively straightforward to measure, it is notoriously difficult to measure accurately and, to compound the problem, it is also extremely variable within a catchment area.

RAINFALL INTENSITY AND STORM DURATION

Water depth is not the only rainfall measure of interest in hydrology; also of importance is the **rainfall intensity** and **storm duration**. These are simple to obtain from an analysis of rainfall records using frequency analysis. The rainfall needs to be recorded at a short time interval (i.e. an hour or less) to provide meaningful data.

Figure 2.16 shows the rainfall intensity for a rain gauge at Bradwell-on-Sea, Essex, UK. It is evident from the diagram that the majority of rain falls at very low intensity: 0.4 mm per hour is considered as light rain. This may be misleading as the rain gauge recorded rainfall every hour and the small amount of rain may have fallen during a shorter period than an hour i.e. a higher intensity but lasting for less than an hour. During the period of measurement there were recorded rainfall intensities greater than 4.4 mm/hr (maximum 6.8 mm/hr) but they were so few as to not show up on the histogram scale used in Figure 2.16. This may be a reflection of only 2 years of records being analysed, which reflects an extremely important concept in hydrology: the **frequency–magnitude** relationship, as introduced in Chapter 1. With rainfall (and runoff – see Chapters 7 and 8) the larger the rainfall event the less frequent we would expect it to be. This is not a linear relationship; as illustrated in Figure 2.16 the curve declines in a non-linear fashion. If we think of the relative frequency as a probability then we can say that the chances of having a low rainfall event are very high: a low magnitude–high frequency event. Conversely the chances of having a rainfall intensity greater than 5 mm/hr are very low (but not impossible): a high magnitude–low frequency event.

In Figure 2.17 the storm duration records for two different sites are compared. The Bradwell-on-Sea site has the majority of its rain events lasting 1 hour or less. In contrast the Ahoskie site has only 20 per cent of its storms lasting 1 hour or less but many more than Bradwell-on-Sea that last 4 hours or more. When the UK site rainfall and intensity curves are looked at together (i.e. Figures 2.16 and 2.17) it can be stated that Bradwell-on-Sea experiences a predominance of low intensity, short duration rainfall events and very few long duration, high intensity storms. This type of information is extremely useful to a hydrologist investigating the likely runoff response that might be expected for the rainfall regime.

In engineering hydrology, extensive use is made of relationships known as depth–duration–frequency (DDF) curves (see Figure 1.12), or the related intensity–duration frequency (IDF) curves. In Figure 2.18, the same data as in Figure 1.12 have been used to derive the curves, the difference being that instead of depth of rainfall (mm), the average intensity (mm/hr) for each duration has been calculated. Whereas in Figure 1.12 the depth increases over time but flattens off, in Figure 2.18, the curve starts off at very high intensity and then decreases over time. In the United Kingdom, the *Flood estimation handbook* (CEH 1999) and accompanying CD-ROM provided access to depth–duration–frequency relationships for any location in the UK (1999 DDF). This has now been superseded by the 2013 DDF relationships available via the FEH Web Service (https://fehweb.ceh.ac.uk). In the US, very useful data are available

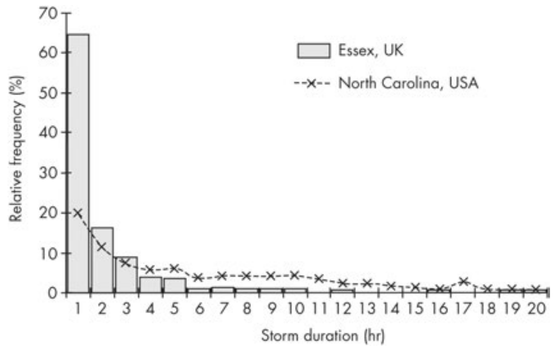

Figure 2.17 Storm duration curves. The bars are for the same data set as Figure 2.16 and the broken line for Ahoskie, North Carolina.

Source: Ahoskie data are redrawn from Wanielista (1990)

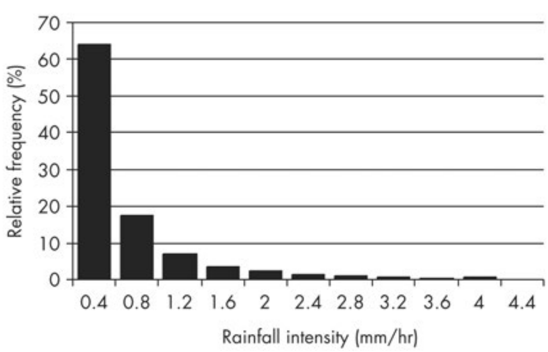

Figure 2.16 Rainfall intensity curve for Bradwell-on-Sea, Essex, UK. Data are hourly recorded rainfall from April 1995 to April 1997.

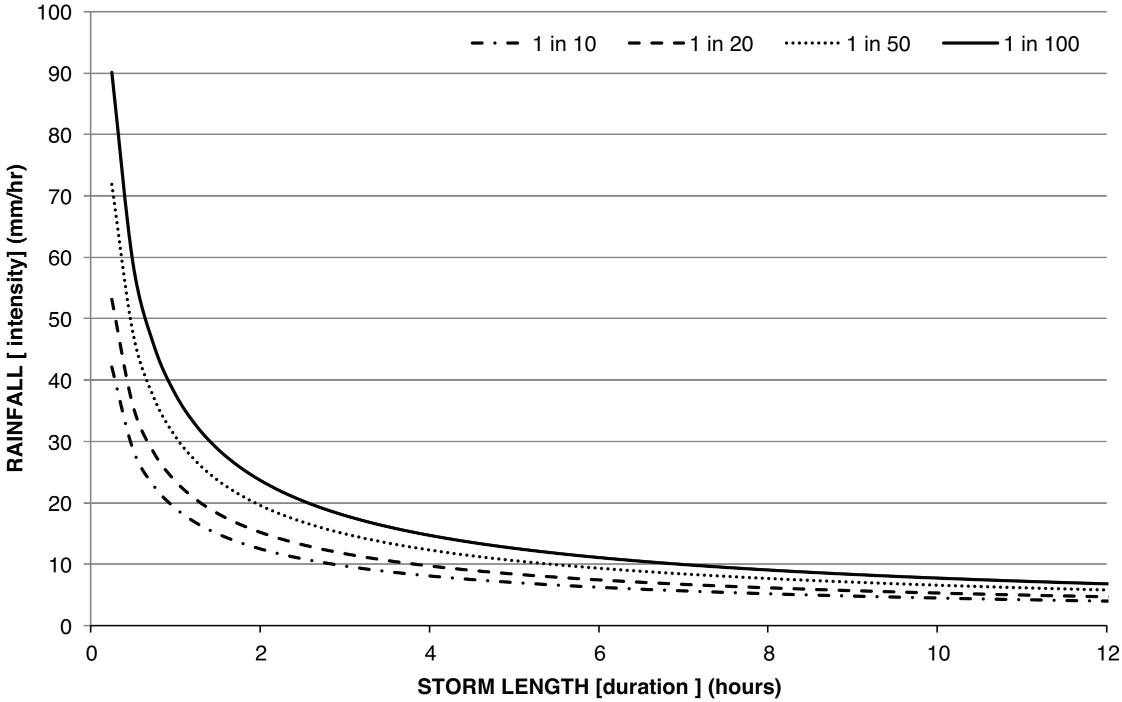

Figure 2.18 Rainfall intensity–duration–frequency (IDF) relationships for the River Boyd catchment, United Kingdom.

Source: Data from CEH (1999)

via NOAA's National Weather Service Precipitation Frequency Data Server (PFDS) (http://hdsc.nws.noaa.gov/hdsc/pfds) which provides State-based depth–duration–frequency data.

SURROGATE MEASURES FOR ESTIMATING RAINFALL

The difficulties in calculation of a spatially distributed precipitation value from point measurements make the direct estimation of areal precipitation an attractive proposition. There are two techniques that make some claim to achieving this: radar, and **satellite remote sensing**. These approaches have many similarities, but they differ fundamentally in the direction of measurement. Radar looks from the earth up into the atmosphere and tries to estimate the amount of precipitation falling over an area. Satellite remote sensing looks from space down towards the earth surface and attempts to estimate the amount of precipitation falling over an area.

Radar

The main use of ground-based radar is in weather forecasting where it is used to track the movement of rain clouds and fronts across the earth's surface. This in itself is interesting but does not provide the hydrological requirement of estimating how much rain is falling over an area.

There are several techniques used for radar, although they are all based on similar principles. Radar is an acronym (*r*adio *d*etection *a*nd *r*anging). A wave of electromagnetic energy is emitted from a unit on the ground and the amount of wave reflection and return time is recorded. The more water there is in a cloud the more electromagnetic energy is reflected back to the ground and detected by the radar unit. The quicker the reflected wave reaches the ground, the closer the cloud is to the surface. The most difficult part of this technique is in finding the best wavelength of electromagnetic radiation to emit and detect. It is important that the electromagnetic wave is reflected by liquid water in the cloud, but not atmospheric gases and/or changing densities of the atmosphere. A considerable amount of research effort has gone into trying to find the best wavelengths for ground-based radar to use. The solution appears to be that it is somewhere in the microwave band (commonly C-band), but that the exact wavelength depends on the individual situation being studied (Cluckie and Collier 1991).

Studies have shown a good correlation between reflected electromagnetic waves and rainfall intensity. Therefore, this can be thought of as a surrogate measure for estimating rainfall. If an accurate estimate of rainfall intensity is required, then a relationship has to be derived using several calibrating rain gauges. Herein lies a major problem with this type of technique: there is no universal relationship that can be used to derive rainfall intensity from cloud reflectivity. An individual calibration has to be derived for each site and this may involve several years of measuring point rainfall coincidentally with cloud reflectivity. This is not a cheap option and the cost prohibits its widespread usage, particularly in areas with poor rain gauge coverage.

In Britain the UK Meteorological Office operates a series of 15 weather radars with a 5 km resolution that provide images every 15 minutes. This is a more intensive coverage than could be expected in most countries. Although portable radar can be used for rainfall estimation, their usage has been limited by the high cost of purchase.

Satellite remote sensing

The atmosphere-down approach of satellite remote sensing is quite different from the ground-up approach of radar – fundamentally because the sensor is looking at the top of a cloud rather than the bottom. It is well established that a cloud most likely to produce rain has an extremely bright and cold top. These are the characteristics that can be observed from space by a satellite sensor. The most common form of satellite sensor is passive (this means it receives radiation from another source, normally the sun, rather than emitting any itself the way radar does) and detects radiation in the visible and infrared wavebands. **LANDSAT**, **SPOT** and **AVHRR** are examples of satellite platforms of this type. By sensing in the visible and infrared part of the electromagnetic spectrum, the cloud brightness (visible) and temperature (thermal infrared) can be detected. This so-called 'brightness temperature' can then be related to rainfall intensity via calibration with point rainfall measurements, in a similar fashion to ground-based radar. One of the problems with this approach is that it is sometimes difficult to distinguish between snow reflecting light on the ground and clouds reflecting light in the atmosphere. They have similar brightness temperature values but need to be differentiated so that accurate rainfall assessment can be made.

Another form of satellite sensor that can be used is passive microwave. The earth emits microwaves (at a low level) that can be detected from space. When there is liquid water between the earth's surface and the satellite sensor (i.e. a cloud in the atmosphere) some of the microwaves are absorbed by the water. A satellite sensor can therefore detect the presence of clouds (or other bodies of water on the surface) as a lack of microwaves reaching the sensor. An early example of a study using a satellite platform that can detect passive microwaves (SSM/I) is in Todd and Bailey (1995), who used the method to assess rainfall over the United Kingdom. Although there was some success in the method it was at a scale of little use to catchment scale hydrology as the best resolution available was around 10 × 10 km grid sizes. More recent satellites (e.g. TRMM and the later GMP (see Case Study below) use on-board K-band microwave transmitters and measure the return of back-scattered microwave radiation.

Although these recent developments have added considerably to our understanding of global precipitation (Figure 2.19 (a) and (b)), their real benefit is in the global scale datasets that are generated,

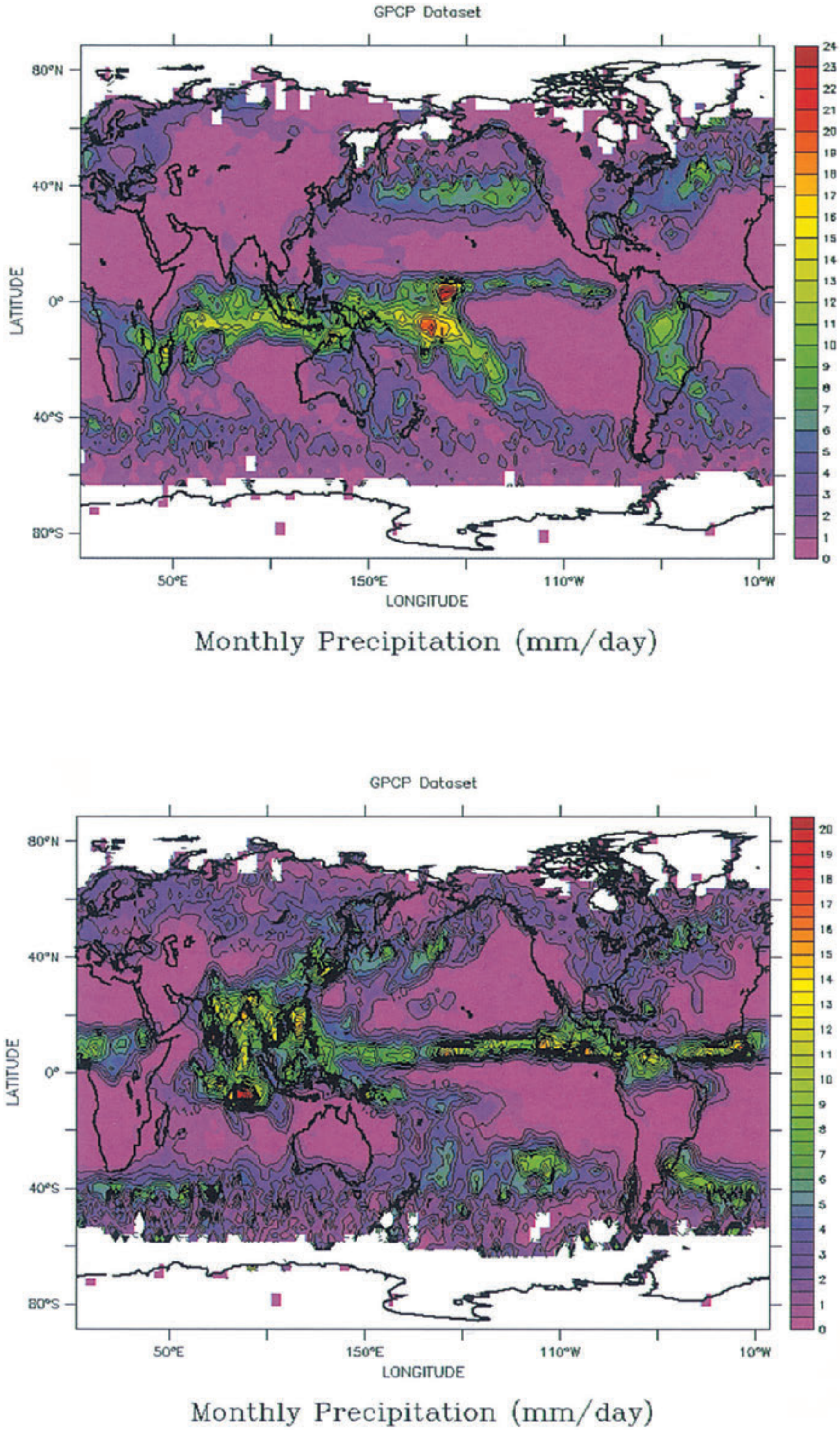

Figure 2.19 Examples of satellite-derived global rainfall distribution in the month of (a) January and (b) July.
Source: Image from NOAA (www.ncdc.noaa.gov)

which can be used in global models. They are also very important in regions where ground measurements are limited (e.g. Kidd et al. 1998), but in countries with high rain gauge density and/or a good ground-based radar network, it does not significantly improve estimation of areal precipitation. It is important to remember too, that satellite remote sensing provides an indirect estimate of precipitation over an area and still requires ground observations for validation.

Case study

ACTIVE REMOTE SENSING OF GLOBAL PRECIPITATION

A commonly accepted limitation of traditional site-based measurement is the lack of continuous, detailed spatio-temporal data. This limits its application in hydrological models. The main benefit of a remote sensing approach to precipitation monitoring is that it is a more cost-effective way to generate representative and continuous rainfall records over larger temporal and spatial scales than was previously possible. Precipitation measurements of this sort can be estimated from the back scattering of microwave radiation emitted from a satellite (this is *active* as opposed to *passive* remote sensing). These active satellites often use K-Band radar enabled instruments to map global scale precipitation at a daily resolution. We will consider two of the most important initiatives in this form of remote sensing.

Tropical Rainfall Measurement Mission (TRMM)

The Tropical Rainfall Measurement Mission (TRMM) was launched in November 1997 as a joint mission between the US National Aeronautics and Space Administration (NASA) and the Japan Aerospace Exploration Agency (JAXA). This was the first satellite dedicated to measuring tropical and subtropical rainfall, and the first satellite to carry specific microwave precipitation radar instrumentation (as well as other sensors). TRMM operated in a non-polar, low inclination orbit (≈400 km above Earth's surface), so focused mainly on the tropical and sub-tropical regions of Earth (35°S–35°N). From this restricted orbital cycle, TRMM completed 16 full cycles per day with a measurement swath of 878 km. The spatial resolution of TRMM data varies between 0.25° and 5°. Although the mission has ended (it was decommissioned in April 2015), it successfully captured high resolution precipitation data for over 15 years, forming an important legacy dataset useful to scientists for detecting and understanding climate variability and change. Products produced by the mission included near real-time 3-hour global rainfall data, 7-day accumulated global rainfall data, and other processed datasets such as averages and anomalies. For more detail on the TRMM mission see: https://pmm.nasa.gov/TRMM.

Global Precipitation Measurement (GPM)

The Global Precipitation Measurement (GPM) mission succeeded TRMM. Although initiated by NASA and JAXA, it is now a wider consortium of several international space agencies, which operates a constellation of satellites. The main satellite, known as the 'GPM Core Observatory' was launched in February 2014. Like the TRMM satellite, it operates in a non-polar, low inclination orbit (407 km above Earth's surface), completing 16 orbits per day – but across a wider proportion of the Earth's surface (65°S–65°N). Using a constellation of satellites enables a revisit time of 1–2 hours. The IMERG algorithm combines data from all passive microwave instruments across the GPM constellation, and provides gridded precipitation datasets at a variety of temporal resolutions (up to 3 hours), and at intervals from close to real time (6-hour lag). The spatial resolution of GPM datasets ranges between 4.4 and 32 km, depending on latitude. For more detail on the GPM mission see: https://pmm.nasa.gov/GPM.

Comparing TRMM & GPM

TRMM, due to its restricted orbit, only monitored precipitation across the global tropics. While it also is non-polar, GPM has a wider monitoring range across middle and high latitudes due to technological advancements in remote sensing instrumentation. GPM also has a higher spatial resolution and improved sensor detection of light rain and snow. Therefore, GPM products are considered to have greater accuracy. Many hydrological systems and applications which used TRMM data are now being reconfigured and are undergoing transition to GPM datasets. These include international extreme rain, flood and drought monitoring systems. Going forward it is likely that TRMM and GPM data products will be inter-calibrated to provide a combined long-term, high temporal resolution, precipitation record.

TRMM & GPM applications

The precipitation datasets produced by TRMM and GPM have been used in a range of hydrometeorological applications.

1. TRMM/GPM has allowed scientists to examine the internal structure of tropical storm systems, increasing our understanding of high rainfall events and implications of such systems beyond the tropics.
2. The microwave radar equipment on-board TRMM, and now GPM, enables scientists to monitor the microscale characteristics of individual raindrops. This allows for the monitoring of the type/intensity/distribution of water in the atmosphere, and increases our understanding of global atmospheric circulation.
3. Beyond water stored in the atmosphere, TRMM/GPM datasets have enabled us to develop understanding around the rate of water/moisture transfer around the hydrological cycle. This often involves combining TRMM/GPM with land surface data.
4. Due to the extensive temporal record of TRMM/GPM datasets scientists can develop an understanding of the changing temporal and spatial extent of extreme floods and droughts at a global scale due to climate change.

Data from both the TRMM and GPM missions is provided by the NASA Goddard Space Flight Centre, free of charge in various formats. To access these datasets go to: https://pmm.nasa.gov/data-access.

The Global Precipitation Climatology Project (GPCP)

Another important international initiative is the Global Precipitation Climatology Project (GPCP) established by the World Climate Research Program (WCRP) in the late 1990s. It holds the longest global precipitation data record (1979–present). The GPCP aims to provide a consistent measure of global precipitation by integrating various satellite and ground-based data sets derived from a mix of observations in space and time.

It provides daily or monthly precipitation at $1° \times 1°$ spatial resolution, from 1979 onwards. It is a core dataset in the analysis of global precipitation and has been used in over 1500 articles in scientific journals. For more information on the GPCP, including access to the datasets, see: http://gpcp.umd.edu.

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PRECIPITATION IN THE CONTEXT OF WATER QUANTITY AND QUALITY

Precipitation, as the principal input into a catchment water balance, has a major part to play in water quantity and quality. By and large, it is the spatial and temporal distribution of precipitation that drives the spatial and temporal distribution of available water. Rainfall intensity frequently controls the amount of runoff during a storm event (see Chapter 7) and the distribution of rain through the year controls the need for irrigation in an agricultural system.

The exception to this is in large river basins where the immediate rainfall distribution may have little bearing on the water flowing down the adjacent river. A good example is the Colorado River which flows through areas of extremely low rainfall in Utah and Arizona. The lack of rainfall in these areas has little bearing on the quantity of water in the Colorado River, it is governed by the precipitation (both rain and snow) falling in the Rocky Mountains well to the north-east.

The influence of precipitation on water quantity directly affects water quality through dilution. Where water quantity is high there is more water available to dilute any contaminants entering a river or groundwater system. It does not follow that high water quantity equates with high water quality but it has the potential to do so.

Precipitation also has a direct influence on water quality through scavenging of airborne pollutants which are then dissolved by the rain. The complex nature of a forest topography means that trees act as recipient surfaces for airborne pollutants. As rain falls onto the tree, salts that have formed on leaves and branches may be dissolved by the water, making the stemflow and throughfall pollutant-rich. This has been observed in field studies, particularly near the edge of tree stands (Neal et al. 1991).

The best known example of pollutant scavenging is **acid rain**. This is where precipitation in areas polluted by industrial smokestacks dissolves gases and absorbs particles that lower the acidity of the rain. Naturally rain is slightly acidic with a pH of between 5 and 6, due to the dissolving of carbon dioxide to form a weak carbonic acid. The burning of fossil fuels adds nitrogen oxides and sulphur oxides to the atmosphere, both of which are easily dissolved to form weak nitric and sulphuric acids. The burning of coal is particularly bad through the amount of sulphur dioxide produced, but any combustion will produce nitrogen oxides by the combination of nitrogen and oxygen (both already in the atmosphere) at high temperatures. In areas of the Eastern United States and Scandinavia rainfall has been recorded with pH values as low as 3 (similar to vinegar). In some situations this makes very little difference to overall water quality as the soil may have enough acid-buffering capability to absorb the acid rain. This is particularly true for limestone areas where the soil is naturally alkaline. However, many soils do not have this buffering capacity due to their underlying geology (e.g. granite areas in the north-east of North America). In this situation the streams become acidic and this has an extremely detrimental effect on the aquatic fauna. The major reason for the impact on fish life is the dissolved aluminium that the acidic water carries; this interferes with the operation of gills and the fish effectively drown.

It is worth noting that the dissolving of nitrous oxide can have a positive effect on plant life through the addition to the soils of nitrate which promotes plant growth (see Chapter 10). To give some scale to the impact of large atmospheric nitrogen inputs, Loÿe-Pilot et al. (1990) estimate that atmospheric nitrogen input into the Mediterranean Sea is of the same order of magnitude as the riverine input. It is estimated that 25 per cent of nitrogen inputs to the Baltic Sea come from the atmosphere (BSEP 2005).

ESSAY QUESTIONS

1. **Describe the different factors affecting the spatial distribution of precipitation at differing scales.**
2. **Give an overview of potential sources of error in the measurement of rainfall and snowfall and explain how they might be minimised?**

3 Compare and contrast different techniques for obtaining a spatially averaged precipitation value (including surrogate measures).

4 Why is scale such an important issue in the analysis of precipitation in hydrology?

FURTHER READING

Bailey, T.C. and Gatrell, A.C. (1995) *Interactive spatial data analysis*. Longman, Harlow.
Gives a modern view of spatial analysis, not necessarily just for precipitation.

Barry, R.G. and Chorley, R. (2003) *Atmosphere, weather and climate* (8th edition). Routledge, London.
Explains the principles of precipitation generation well.

Shaw, E.M., Bevan, K.J., Chappell, N.A. and Lamb, R. (2011) *Hydrology in practice* (4th edition). Spon Press, London.
'Chapter 3: Precipitation' gives a good overview of both the history of precipitation measurement and current UK practice, while 'Chapter 9: Precipitation analysis' reviews analytical approaches.

Strangeways, I. (2006) *Precipitation: Theory, measurement and distribution*. Cambridge University Press, Cambridge.
A modern text on precipitation processes and measurement.

Sumner, G. (1988) *Precipitation: Process and analysis*. John Wiley & Sons, Chichester.
Somewhat dated, but nevertheless a classic text on precipitation.

3

EVAPORATION

LEARNING OBJECTIVES

When you have finished reading this chapter you should have:

- An understanding of the fundamental process of evaporation and what controls its rate.
- A knowledge of the techniques for measuring evaporation directly.
- A knowledge of the techniques used to estimate evaporation.
- An understanding of the importance of vegetation in determining regional evaporation.

Evaporation is the transferral of liquid water into a gaseous state and its diffusion into the atmosphere (Figure 1.4). In order for this to occur there must be liquid water present and available energy from the sun or atmosphere. The importance of evaporation within the hydrological cycle depends very much on the amount of water present and the available energy, two factors determined by a region's climate. During winter months in humid-temperate climates, evaporation may be a minor component of the hydrological cycle as there is very little available energy to drive the evaporative process. This alters during summer when there is abundant available energy and evaporation has the potential to become a major part of the water balance. The potential may be limited by the availability of liquid water during the dry months. This can be seen in extremely hot, arid climates where there is often plenty of available energy to drive evaporation but very little water to be evaporated. As a consequence, the actual amount of evaporation is small.

It is the presence or lack of water at the surface that provides the major semantic distinction in definitions of the evaporative process. **Open water evaporation** (often denoted as E_o) is the evaporation that occurs above a body of water such as a lake, stream or the oceans. Figure 1.7 shows that at the global scale this is the largest source of evaporation, in particular from the oceans. **Potential evaporation** (*PE*) is that which occurs over the land's surface, or would occur if the water supply were unrestricted. This occurs when a soil is wet and what evaporation is able to happen occurs without a lack of water supply. **Actual evaporation** (E_t) is that which actually occurs (i.e. if there is not much available water it will be less than potential). When conditions are very wet (e.g. during a rainfall event) E_t will equal PE, otherwise it will be less than PE. In hydrology we are most interested in E_o and E_t but normally require PE to calculate the E_t value.

All of these definitions have been concerned with 'evaporation over a surface'. In hydrology the surface is either water (river, lake, ponds, etc.) or the land. The evaporation above a land surface occurs in two ways – either as actual evaporation from the soil matrix or **transpiration** from plants. The combination of these two is often referred to as **evapotranspiration**, although the term *actual evaporation* is essentially the same (hence the t subscript in E_t). Transpiration from a plant occurs as part of photosynthesis and respiration. The rate of transpiration is controlled by the opening or closing of stomata in the leaf. Transpiration can be ascertained at the individual plant level by instruments measuring the flow of water up the trunk or stem of a plant. Different species of plants transpire at different rates but the fundamental controls are the available water in the soil, the plant's ability to transfer water from the soil to its leaves and regulate its stomata, and the ability of the atmosphere to absorb the transpired water.

Evaporation is sometimes erroneously described as the only loss within the water balance equation. The water balance equation is a mathematical description of the hydrological cycle and by definition there are no losses and gains within this cycle. What is meant by 'loss' is that evaporation is lost from the earth's surface, since hydrologists are mostly concerned with the water that contributes to sustaining streamflow. To a meteorologist concerned with the atmosphere, evaporation can be seen as a gain. Evaporation, although not a loss, can be viewed as the opposite of precipitation, particularly in the case of dewfall, a form of precipitation. In this case the **dewfall** (or negative evaporation) is a gain to the earth's surface.

EVAPORATION AS A PROCESS

We will begin by taking a closer look at the process of evaporation. As mentioned in the introduction, evaporation means the transferral of liquid water into a gaseous state. In liquid, water molecules are moving constantly (*Brownian motion*) and adding additional heat (energy) causes water molecules to move more rapidly and further apart, weakening the forces holding them together. At higher temperatures, some water molecules near the surface have sufficient energy to break free from the surface of the water and escape into the surrounding air. Think of this as the 'boiling-off rate' or rate of *vaporisation* (Figure 3.1).

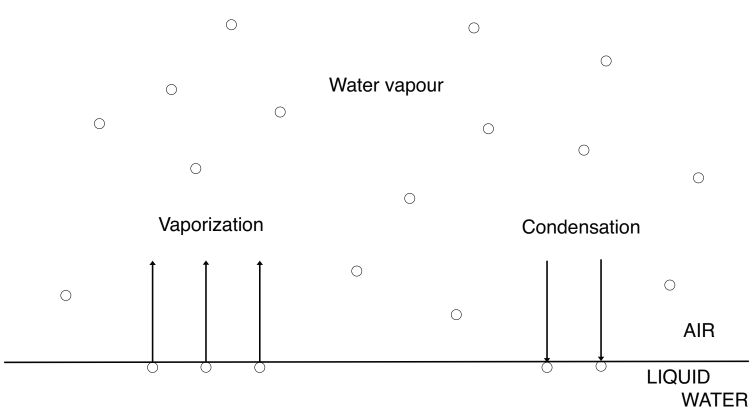

Figure 3.1 Evaporation is the net balance between the rate of vaporisation of water molecules into the atmosphere and the condensation of water molecules from the atmosphere into liquid water.

Source: Redrawn from Ward and Robinson (2000)

Even the driest air contains some water vapour and these molecules too, are in constant motion. Some of those close to the water surface will move into the liquid water – think of this as the 're-entry' rate, or rate of *condensation*. Evaporation is the net balance between the two processes of vaporisation and condensation (Ward and Robinson 2000).

Now think of a half-filled, closed bottle of water standing in the sun. Liquid water will evaporate, and this is controlled by the temperature of the liquid water. As water vaporises, the concentration of water molecules in the top half increases, as does the vapour pressure. The vapour pressure in turn controls the condensation rate: the higher the vapour pressure, the higher the condensation rate. Since in a sealed container the vapour has nowhere to go, the rates of vaporisation and condensation eventually equal each other, and this is the point at which the air is *saturated*.

It has already been said that evaporation requires an energy source and an available water supply to transform liquid water into water vapour. There is one more precondition: that the atmosphere be dry enough to receive any water vapour produced (as we have seen above, evaporation ceases once the air becomes saturated). These are the three fundamental parts to an understanding of the evaporation process. This was first understood by John Dalton (1766–1844), an English physicist who linked wind speed and the dryness of the air to the evaporation rate.

Available energy

The main source of energy for evaporation is from the sun. This is not necessarily in the form of direct radiation, it is often absorbed by a surface and then re-radiated at a different wavelength. The normal term used to describe the amount of energy received at a surface is **net radiation** (Q^*), measured using a net radiometer. Net radiation is a sum of all the different heat fluxes found at a surface and can be described by Equation 3.1.

$$Q^* = Q_S \pm Q_L \pm Q_G \tag{3.1}$$

where Q_S is the sensible heat flux; Q_L is the latent heat flux and Q_G is the soil heat flux.

Sensible heat is that which can be sensed by instruments. This is most easily understood as the heat we feel as warmth. The sensible heat flux is the rate of flow of that sensible heat. **Latent heat** is the heat either absorbed or released during a phase change from ice to liquid water, or liquid water to water vapour. When water moves from liquid to gas this is a negative flux (i.e. energy is absorbed) whereas the opposite phase change (gas to liquid) produces a positive heat flux.

The **soil heat flux** is heat released from the soil having been previously stored within the soil. This is frequently ignored as it tends to zero over a 24-hour period and is a relatively minor contributor to net radiation.

Incoming solar radiation is filtered by the atmosphere so that not all the wavelengths of the electromagnetic spectrum are received at the earth's surface. Incoming radiation that reaches the surface is often referred to as short-wave radiation: visible light plus some bands of the infrared. This is not strictly true as clouds and water vapour in the atmosphere, plus trees and tall buildings above the surface, emit longer-wave radiation which also reaches the surface.

Outgoing radiation can be either reflected short-wave radiation or energy radiated back by the earth's surface. In the latter case this is normally in the infrared band and longer wavelengths and is referred to as long-wave radiation. This is a major source of energy for evaporation.

There are two other forms of available energy that under certain circumstances may be important sources in the evaporation process. The first is heat stored in buildings from an anthropogenic source (e.g. domestic heating). This energy source is often fuelled from organic sources and may be a significant addition to the heat budget in an urban environment, particularly in the winter months. The second additional source is **advective energy**. This is energy that originates from elsewhere (another region that may be hundreds or thousands of kilometres away) and has been transported to the evaporative surface (frequently in the form of latent heat) where it becomes available energy in the form of sensible heat. The best example of this is latent energy that arrives in cyclonic storm systems. In Chapter 1 it was explained that evaporating and condensing water is a major means of redistributing energy around the globe. The evaporation of water that contributes to cyclonic storms normally takes place over an ocean, whereas the condensation may occur a considerable distance away. At the time of evaporation, thermal energy (i.e. sensible heat) is transferred into latent energy that is then carried by the water vapour to the place of condensation where it is released as sensible heat once more. This 're-release' is often referred to as advective energy and may be a large energy source to drive further evaporation.

Water supply

As we saw in Chapter 1, evaporation from the ocean is a major component of the hydrological cycle. However, in terrestrial environments, the available water supply can be from water directly on the surface of a lake, river or pond. In this case it is open water evaporation (E_o). When the water is lying within soil the water supply becomes more complex. Soil water can evaporate directly, although it is normally only from the near surface. As the water is removed from the surface it sets up a soil moisture gradient that will draw water from deeper in the soil towards the surface, but it must overcome the force of gravity and the withholding force exerted by soil capillaries (see Chapter 6). In addition to this the water may be brought to the surface by plants using osmosis in their rooting system. The way that soil moisture controls the transformation from potential evaporation to actual evaporation is complex and will be discussed further later in this chapter.

The receiving atmosphere

Once the available water has been transformed into water vapour, using whatever energy source is available, it then must be absorbed into the atmosphere surrounding the surface. This process of *diffusion* requires that the atmosphere is not already saturated with water vapour and that there is enough buoyancy to move the water vapour away from the surface. These two elements can be assessed in terms of the **vapour pressure deficit** and atmospheric mixing.

In Chapter 2 we explored the relationship between temperature and vapour pressure, highlighting that the maximum amount of water vapour that can be held in a parcel of air is defined by the curve in Figures 2.1 and 3.2. At all the points along the curve, the air is saturated or at absolute humidity. You will recall from Chapter 2 that absolute humidity is the maximum mass of water that air can obtain (measured in g/kg) at each of the temperatures along the curve. However, let us assume that the atmosphere is not saturated at a particular temperature, but located at point A in Figure 3.2 (actual vapour pressure = 10.0 hPa). The **relative humidity** of the atmosphere (i.e. how close to fully saturated the atmosphere is) is defined as the ratio of the actual vapour pressure relative to the maximum possible at that temperature (saturated vapour pressure = 23.4 hPa), expressed as a percentage. So at point A, the relative humidity is 42.7 per cent (10.0 hPa/23.4 hPa × 100). The difference between the actual vapour pressure (point A) and the saturation vapour pressure (point B) is the *vapour pressure deficit* (vpd), also called the *saturation deficit*. The vpd is a measure of how much extra water vapour the atmosphere could hold assuming a constant temperature and pressure. In Figure 3.2 the vpd is 13.4 hPa (23.4 – 10 = 13.4 hPa), so that air is capable of holding enough extra water vapour to create an additional 13.4 hPa of vapour pressure. The higher the vpd the more water can be absorbed from an evaporative surface.

Atmospheric mixing is a general term meaning how well a parcel of air is able to diffuse into the atmosphere surrounding it. The best indicator of atmospheric mixing is the wind speed at different heights above an evaporating surface. If the wind speed is zero, the parcel of air will not move away from the evaporative surface and will 'fill' with water vapour. As the wind speed increases, the parcel of air will be moved quickly on to be replaced by another, possibly drier, parcel ready to absorb more water vapour. If the evaporative surface is large (e.g. a lake) it is important that

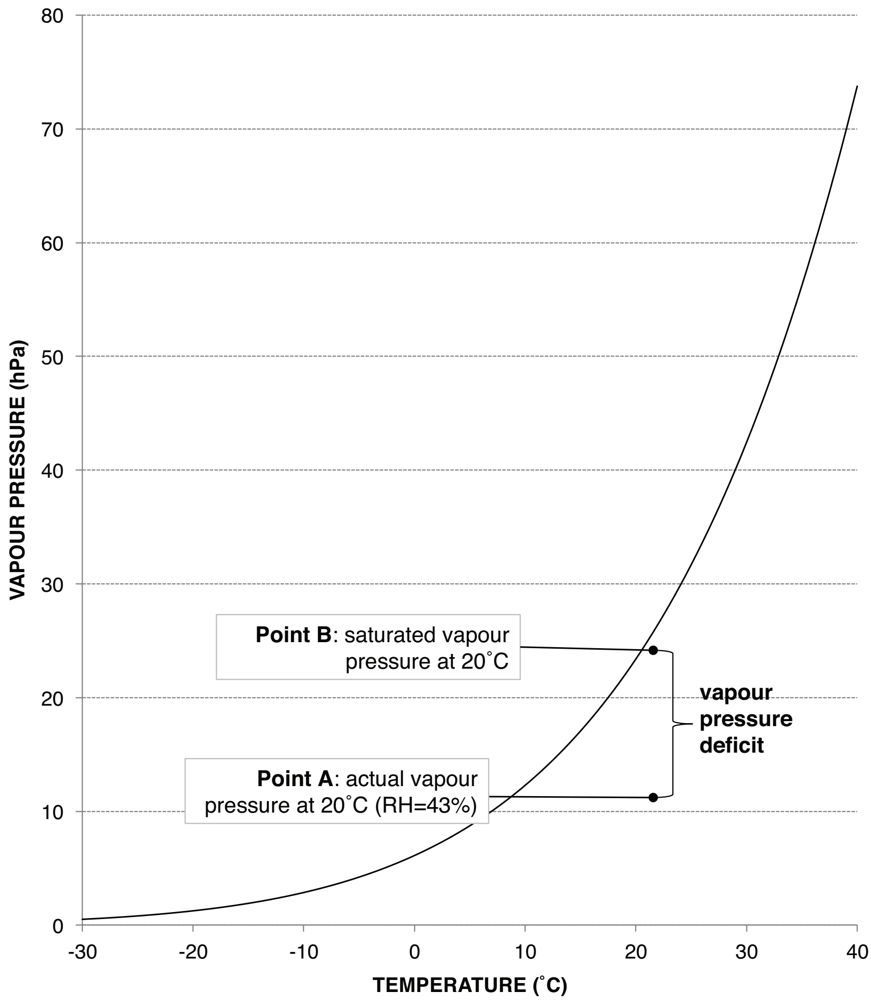

Figure 3.2 The difference between the *actual vapour pressure* at 20 °C and the *vapour pressure at saturation* defines the vapour pressure deficit.

the parcel of air moves up into the atmosphere, rather than directly along at the same level, so that there is drier air replacing it. This occurs through turbulent diffusion of the air. There is a greater turbulence associated with air passing over a rough surface than a smooth one, something that will be returned to in the discussion of evaporation estimation.

One way of thinking about evaporation is in terms of a washing line. The best conditions for drying your washing outside are on a warm, dry and windy day. Under these circumstances the evaporation from your washing (the available water) is high due to the available energy being high (it is a warm day), and the receiving atmosphere mixes well (it is windy) and is able to absorb much water vapour (the air is dry). On a warm and still day, or a warm and humid day washing does not dry as well (i.e. the evaporation rate is low). Understanding evaporation in these terms allows us to think about what the evaporation rate might be for particular atmospheric conditions.

Open water evaporation

As is clear from the earlier part of this chapter, open water evaporation (E_o), is the most straightforward case of evaporation. This includes water evaporating from lakes and rivers, but also water intercepted by vegetation, buildings and other surfaces. We will deal with intercepted water and its evaporation in the next chapter, and will concentrate on evaporation from water bodies in this section.

Thinking back to the fundamental processes discussed at the start of the chapter, it is the temperature of the surface layers of the water body that will control the vaporisation rate (properties of the water body). The condensation rate is controlled by the vapour pressure of the layer of air above the water body, in other words the properties of the receiving atmosphere. This means the vapour pressure deficit, but also the atmospheric mixing as described in the previous section. Evaporation loss from water bodies can be significant, particularly in arid environments. Properties of the water body that are important include the surface area, depth and salinity (evaporation decreases by 1 per cent for every 1 per cent increase in salinity) (Ward and Robinson 2000).

Soil evaporation

Although the fundamental process and drivers of evaporation from soils are exactly the same as evaporation from open water, evaporation from soil is usually less than that from open water, under the same meteorological conditions. There are a number of reasons for this. Firstly, the supply of water is more limited relative to open water. Secondly, for a supply of moisture to be available at the surface, it needs to be transmitted upwards from lower down in the soil profile. As we shall see in Chapter 6, properties of the soil exert strong controls on the movement of water, particularly if the soil is not saturated. Evaporation directly from the soil is particularly relevant when considering bare earth, such as ploughed land.

Transpiration and total evaporation

Transpiration by a plant leads to evaporation from leaves through small holes (stomata) in the leaf. This is sometimes referred to as dry leaf evaporation. The influence of stomata on the transpiration rate is an interesting plant physiological phenomenon. Some plants are very effective at shutting stomata when under water stress, and therefore limit their water usage. The water stress occurs when the vapour pressure deficit is high and there is a high evaporative demand. In this situation the stomata within a leaf can be likened to a straw. When you suck hard on a soft straw it creates a pressure differential between the inside and outside and the sides collapse in; therefore you cannot draw air easily through the straw. Stomata can act in a similar manner so that when the evaporative demand is high (sucking water vapour through the stomata) the stomata close down and the transpiration rate decreases. Some plant species shut their stomata when under evaporative stress (e.g. conifers) while others continue transpiring at high rates when the evaporative demand is high (e.g. many pasture species). The ability of plants to shut their stomata can influence the overall water budget as their overall evaporation is low. This is illustrated in the case study later in this chapter on using a lysimeter to measure tussock evaporation.

Where there is a vegetation canopy the evaporation above this surface will be a mixture of transpiration, evaporation from the soil and evaporation from wet leaves (*canopy interception* or *interception loss* or *wet leaf evaporation* – see Chapter 4). The relative importance of these three evaporation sources will depend on the degree of vegetation cover and the climate at the site. In tropical rainforests transpiration is the dominant water loss but where there is a seasonal soil water deficit the influence of canopy interception loss becomes more important. This is illustrated by the data in Table 3.1 which contrasts the water balance for two *Pinus radiata* forests at different locations in New Zealand (with different climates).

Wetland environments are interesting in that there is both open water evaporation and transpiration occurring simultaneously. In humid conditions (moist air, no wind) atmospheric demand would be very low because of the high moisture content of the air. However, when the air is dry and/or the wind is blowing dry air in, open water evaporation will be occurring at a high rate, as well as transpiration.

Table 3.1 Estimated evaporation losses from two *Pinus radiata* sites in New Zealand

	Puruki (Central North Island, NZ) (% annual rainfall in brackets)		*Balmoral (Central South Island, NZ) (% annual rainfall in brackets)*	
Annual rainfall	1,405mm		870mm	
Annual interception loss	370mm	(26)	220mm	(25)
Annual transpiration	705mm	(50)	255mm	(29)
Annual soil evaporation	95mm	(7)	210mm	(24)
Remainder (runoff + percolation)	235mm	(17)	185mm	(21)

Source: Data adapted from Kelliher and Jackson (2001)

MEASUREMENT OF EVAPORATION

In the previous chapter there was much emphasis on the difficulties of measuring precipitation due to its inherent variability. All these difficulties also apply to the measurement of evaporation, but they pale into insignificance when you consider that now we are dealing with measuring the rate at which a gas (water vapour) moves away from a surface. Concentrations of gases in the atmosphere are difficult to measure, and certainly there is no gauge that we can use to measure total amounts in the same way that we can for precipitation.

In each of the process chapters in this book there is an attempt to distinguish between measurement and estimation techniques. In the case of evaporation this distinction becomes extremely blurred. In reality almost all the techniques used to find an evaporation rate are estimates, but some are closer to true measurement than others. In this section each technique will include a sub-section on how close to 'true measurement' it is.

Direct micro-meteorological measurement

There are three main methods used to measure evaporation directly: the eddy fluctuation (or correlation), aerodynamic profile, and **Bowen ratio** methods. These are all micro-meteorological measurement techniques and details on them can be found elsewhere (e.g. Oke 1987). An important point to remember about them all is that they are attempting to measure how much water is being evaporated above a surface – a very difficult task.

The eddy fluctuation method measures the water vapour above a surface in conjunction with a vertical wind speed and temperature profiles. These have to be measured at extremely short timescales (e.g. microseconds) to account for eddies in vertical wind motion. Consequently, extremely detailed micro-meteorological instrumentation is required with all instruments having a rapid response time. In recent years this has become possible with hot wire **anemometers** and extremely fine thermistor heads for thermometers. One difficulty is that you are necessarily measuring over a very small surface area and it may be difficult to scale up to something of interest to catchment-scale hydrology.

The aerodynamic profile (or turbulent transfer) method is based on a detailed knowledge of the energy balance over a surface. The fundamental idea is that by calculating the amount of energy available for evaporation, the actual evaporation rate can be determined. The measurements required are changes in temperature and humidity giving vertical humidity gradients. To use this method it must be assumed that the atmosphere is neutral and stable, two conditions that are not always applicable.

The Bowen ratio method is similar to the aerodynamic profile method but does not assume as much about the atmospheric conditions. The Bowen ratio is the ratio of sensible heat to latent heat and requires detailed measurement of net radiation, soil heat flux, temperature and humidity gradient above a surface. These measurements need to be averaged over a 30-minute period to allow the inherent assumptions to apply.

All of these micro-meteorological approaches to measuring evaporation use sophisticated instruments that are difficult to leave in the open for long periods of time. In addition to this they are restricted in their spatial scope (i.e. they only measure over a small area). With these difficulties it is not surprising that they tend to be used at a very small scale, mostly to calibrate estimation techniques (see pp. 58–65). They are accurate in the assessment of an evaporation rate, hence their use as a standard for the calibration of estimation techniques. The real problem for hydrology is that it is not a robust method that can be relied on for long periods of time.

Indirect measurement (water balance techniques)

Evaporation pans

The most common method for the measurement of evaporation is using an **evaporation pan** (see Figure 3.3). This is a large pan of water with a water depth measuring instrument or weighing device underneath that allows you to record how much water is lost through evaporation over a time period. This technique is actually a manipulation of the water balance equation, hence the terminology used here of a water balance technique. An evaporation pan is constructed from impervious material and the water level is maintained below the top so that no seepage or leakage occurs. This eliminates runoff (Q term) from the water balance. Therefore it can be assumed that any change in storage is related to either evaporative loss or precipitation gain. This means that the water balance equation can be rearranged as shown in Equation 3.2:

$$E = \Delta S - P \tag{3.2}$$

If there is a precipitation gauge immediately adjacent to the evaporation pan then the P term can be accounted for, leaving only the change in storage (ΔS) to be measured as either a weight loss or a drop in water depth. At a standard meteorological station the evaporation is measured daily as the change in water depth. For a finer temporal resolution (e.g. hourly) there are load cell instruments available which measure and record the weight at regular intervals.

An evaporation pan is filled with water, hence you are measuring E_o, the open water evaporation. Although this is useful, there are severe problems with using this value as an indicator of actual evaporation (E_t) in a catchment. The first problem is that E_0 will normally be considerably higher than E_t because the majority of evaporation in a catchment will be occurring over a land surface where the available water is contained within soil and may be limited. This will lead to a large overestimation of the actual evaporation. This factor is well known and consequently evaporation pans are rarely used in catchment water balance studies, although they are useful for estimating water losses from lakes and reservoirs.

There are also problems with evaporation pans that make them problematic even for open water evaporation estimates. A standard evaporation pan, called a Class A evaporation pan, is 1,207 mm in diameter and 254 mm deep. The size of the pan makes them prone to the 'edge effect'. As warm air blows across a body of water it absorbs any water vapour evaporated from the surface. Numerous studies have shown that the evaporation rate is far higher near the edge of the water than towards the centre where the air is able to absorb less water vapour (this also applies to land surfaces). The small size of an evaporation pan means that the whole pan is effectively an 'edge' and will

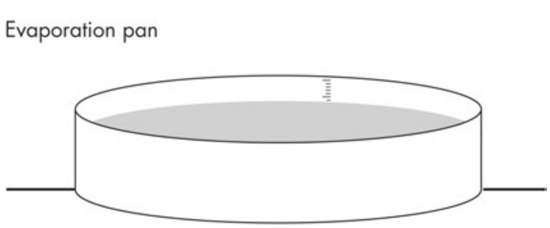

Figure 3.3 An evaporation pan. This sits above the surface (to lessen rain splash) and has either an instrument to record water depth or a continuous weighing device, to measure changes in volume.

have a higher evaporation rate than a much larger body of water. A second, smaller, problem with evaporation pans is that the sides, and the water inside, will absorb radiation and warm up quicker than in a much larger lake, providing an extra energy source and greater evaporation rate.

To overcome the edge effect, empirical (i.e. derived from measurement) coefficients can be used which link the evaporation pan estimates to larger water body estimates. Doorenbos and Pruitt (1975) give estimates for these coefficients that require extra information on upwind fetch distance, wind run and relative humidity at the pan (Goudie et al. 1994). Grismer et al. (2002) provide empirical relationships linking pan evaporation measurements to potential evapotranspiration, i.e. from a vegetated surface, not open water evaporation.

Lysimeters

A **lysimeter** takes the same approach to measurement as the evaporation pan, the fundamental difference being that a lysimeter is filled with soil and vegetation as opposed to water (see Figure 3.4). This difference is important, as E_t rather than E_o is being indirectly measured. A lysimeter can also be made to blend in with the surrounding land cover, lessening the edge effect described for an evaporation pan.

There are many versions of lysimeters in use, but all use some variation of the water balance equation to estimate what the evaporation loss has been. One major difference from an evaporation pan is that a lysimeter allows percolation through the bottom, although the amount is measured. Percolation is necessary so that the lysimeter mimics as closely as possible the soil surrounding it; without any it would fill up with water. In the same manner as an evaporation pan it is necessary to measure the precipitation input immediately adjacent to the lysimeter. Assuming that the only runoff (Q) is through percolation, the water balance equation for a lysimeter is shown in Equation 3.3.

$$E = \Delta S - P - Q \qquad (3.3)$$

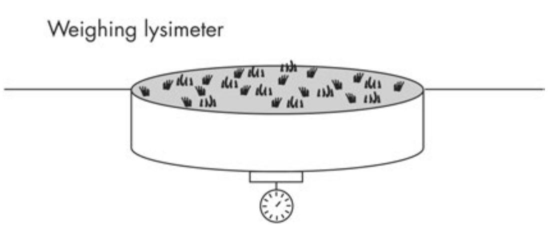

Figure 3.4 A weighing lysimeter sitting flush with the surface. The cylinder is filled with soil and vegetation similar to the surroundings.

A lysimeter faces similar problems to a rain gauge in that it is attempting to measure the evaporation that would be lost from a surface if the lysimeter were not there. The difference from a rain gauge is that what is contained in the lysimeter should closely match the surrounding plants and soil. Although it is never possible to recreate the soil and plants within a lysimeter perfectly, a close approximation can be made and this represents the best efforts possible to measure evaporation. Although lysimeters potentially suffer from the same edge effect as evaporation pans, the ability to match the surrounding vegetation means there is much less of an edge effect.

A *weighing lysimeter* has a weighing device underneath that allows any change in storage to be monitored. This can be an extremely sophisticated device (e.g. Campbell and Murray 1990; Yang et al. 2000); where percolation is measured continuously using the same mechanism for a tipping-bucket rain gauge, weight changes are recorded continuously using a hydraulic pressure gauge, and precipitation is measured simultaneously. A variation on this is to have a series of small weighing lysimeters (such as small buckets) that can be removed and weighed individually every day to provide a record of weight loss. At the same time as weighing, the amount of percolation needs to be recorded. This is a very cheap way of estimating evaporation loss for a study using low technology.

Without any instrument to weigh the lysimeter (this is sometimes referred to as a *percolation gauge*) it must be assumed that the change in soil moisture over a period is zero and therefore evaporation equals rainfall minus runoff. This may be a reasonable assumption over a long time period such as a year where the soil storage will be approximately the same between two winters. An example of this type of lysimeter was the work of Law who investigated the effect that trees had on the water balance at Stocks Reservoir in Lancashire, UK (Law 1956; see Case Study on p. 74).

A well-planned and executed lysimeter study probably provides the best information on evaporation that a hydrologist could find. However, it must be remembered that it is not evaporation that is being measured in a lysimeter – it is almost everything else in the water balance equation, with an assumption being made that whatever is left must be caused by evaporation. One result of this is that any errors in measurement of precipitation and/or percolation will transfer and possibly magnify into errors of evaporation measurement.

ESTIMATION OF EVAPORATION

The difficulties in measuring evaporation using either micro-meteorological instruments (problematic when used over long time periods and at the catchment scale) or water balance techniques (accumulated errors and small scale) have led to much effort being placed on estimating evaporation rather than trying to actually measure it. Some of the techniques outlined below are complicated and this sometimes leads hydrologists to believe that they are measuring, rather than estimating, evaporation. What they are actually doing is taking climatological variables that are known to influence evaporation and simulating evaporation rates from these: an estimation technique. The majority of research efforts in this field have been to produce models to estimate evaporation; however, more recently, satellite remote sensing has provided another method of estimating the evaporation flux.

The techniques described here represent a range of sophistication and they are certainly not all universally applicable. Almost all of these are concerned with estimating the potential evaporation over a land surface. As with most estimation techniques, the hydrologist is required to choose the best techniques for the study situation. In order to help in this decision the various advantages and shortcomings of each technique are discussed.

Thornthwaite

Thornthwaite derived an empirical model (i.e. derived from measurement not theoretical understanding) linking average air temperature to potential evaporation. This is an inherently sensible link in that we know air temperature is closely linked to both available energy and the ability of air to absorb water vapour.

The first part of the Thornthwaite estimation technique (Thornthwaite 1944, 1954) derives a monthly heat index (i) for a region based on the average temperature t (°C) for a month (Equation 3.4).

$$i = \left(\frac{t}{5}\right)^{1.514} \tag{3.4}$$

These terms are then summed to provide an annual heat index I (Equation 3.5).

$$I = \sum_{j=1}^{12} i \tag{3.5}$$

Thornthwaite then derived an equation to provide evaporation estimates based on a series of observed evaporation measurements (Equation 3.6).

$$PE = 16b\left(\frac{10t}{I}\right)^{a} \tag{3.6}$$

The a and b terms in this equation can be derived in the following ways. Term b is a correction factor to account for unequal day length between months. Its value can be found by looking up tables based on the latitude of your study site. Term a is calibrated as a cubic function from the I term such as is shown in Equation 3.7.

$$a = 6.7 \times 10^{-7} I^3 - 7.7 \times 10^{-5} I^2 + 0.018\, I + 0.49 \tag{3.7}$$

Case study

A LYSIMETER USED TO MEASURE EVAPORATION FROM TUSSOCK

A narrow-leafed tussock grass (*Chionochloa rigida*, commonly called 'snow' or 'tall tussock') covers large areas of the South Island of New Zealand. A field study of a catchment dominated by snow tussock (Pearce et al. 1984) showed high levels of baseflow (i.e. high levels of streamflow between storm actions). Mark et al. (1980) used a percolation gauge under a single tussock plant and estimating evaporation, showed that the water balance can show a surplus. They suggested that this may be due to the tussock intercepting fog droplets that are not recorded as rainfall in a standard rain gauge (Figure 3.5). The nature of a tussock leaf (long and narrow with a sharp point), would seem to be conducive to fog interception in the same manner as conifers intercept fog. Another interpretation of the Mark et al. (1980) study is that the estimation of evaporation was incorrect. An understanding of the mechanisms leading to high baseflow levels is important for a greater understanding of hydrological processes leading to streamflow.

In order to investigate this further, a large lysimeter was set up in two different locations. The lysimeter was 2 m in diameter and contained nine mature snow tussock plants in an undisturbed monolith, weighing approximately 8,000 kg. Percolating runoff was measured with a tipping-bucket mechanism and the whole lysimeter was on a beam balance giving a sensitivity of 0.054 mm (Figure 3.6). The rainfall was measured immediately adjacent to the lysimeter. Campbell and Murray (1990) show that although there were times when fog interception appeared to occur (i.e. the catch in the lysimeter was greater than that in the nearby rain gauge) this only accounted for 1 per cent of the total precipitation. The detailed micro-meteorological measurements showed that the tussock stomatal or canopy resistance term was very high and that the plants had an ability to stop transpiring when the water stress became too high (see earlier discussion on plant physiological response to evaporative stress). The conclusion from the study was that snow tussocks are conservative in their use of water, which would appear to account for the high baseflow levels from tussock-covered catchments (Davie et al. 2006).

Figure 3.5 Water droplets condensing on the end of tussock leaves during a fog.

Figure 3.6 Large weighing lysimeter at Glendhu being installed. The weighing mechanism can be seen underneath.

Photo: Barry Fahey

The Thornthwaite technique is extremely useful as potential evaporation can be derived from knowledge of average temperature (often readily available from nearby weather stations) and latitude. There are drawbacks to its usage however; most notably that it only provides estimates of monthly evaporation. For anything at a smaller time-scale it is necessary to use another technique such as Penman's (see below). There are also problems with using Thornthwaite's model in areas of high potential evaporation. The empirical nature of the model means that it has been calibrated for a certain set of conditions and that it may not be applicable outside these. The Thornthwaite model has been shown to underestimate potential evaporation in arid and semi-arid regions (e.g. Acheampong 1986). If the model is being applied in conditions different to Thornthwaite's original calibration (humid temperate regions) it is advisable to find out if any researcher has published different calibration curves for the climate in question.

Penman

Penman was a British physicist who derived a theoretical model of evaporation. Penman's first theoretical model was for open water evaporation and is shown in Equations 3.8 and 3.9 (Penman 1948):

$$E_o = \frac{\Delta Q^* + \gamma E_a}{\Delta + \gamma} \tag{3.8}$$

where an empirical relationship states that:

$$E_a = 2.6\delta_e\left(1 + \frac{u}{1.862}\right) \tag{3.9}$$

and Q^* = net radiation (in evaporation equivalent units of mm/day)
Δ = rate of increase of the saturation vapour deficit with temperature (kPa/°C see Figure 3.6)
δ_e = vapour pressure deficit of the air (kPa)
γ = psychrometric constant (≈ 0.063 kPa/°C)
u = wind speed at 2 m elevation (m/s)

In his original formula, Penman estimated net radiation from empirical estimates of short- and long-wave radiation. The formula given here requires observations of temperature, wind speed, vapour pressure (which can be derived from relative humidity) and net radiation and gives the evaporation in units of mm per day. All of these can be obtained from meteorological measurement. It is normal to use daily averages for these variables, although Shuttleworth (1988) has suggested that it should not be used for time steps of less than 10 days. There are several different ways of presenting this formula, which makes it difficult to interpret between texts. The main difference is in whether the evaporation is a flux or an absolute rate. In the equation above, terms like 'net radiation' have been divided by the amount of energy required to evaporate 1 mm of water (density of water (ρ) multiplied by the latent heat of vaporisation (λ)) to turn them into water equivalents. This means the equation derives an absolute value for evaporation rather than a flux.

Penman continued his work to consider the evaporation occurring over a vegetated surface (Penman and Scholfield 1951), while others refined the work (e.g. van Bavel 1966). Part of this refinement was to include a term for aerodynamic resistance (r_a) to replace E_a (Equation 3.9). **Aerodynamic resistance** is a term to account for the way in which the water evaporating off a surface mixes with a potentially drier atmosphere above it through turbulent mixing. The rougher the canopy surface the greater degree of turbulent mixing that will occur since air passing over the surface is buffeted around by protruding objects. As it is a resistance term, the higher the value, the greater the resistance to mixing; therefore a forest has a lower value of r_a than smoother pasture. Some values of aerodynamic resistance for different vegetation types are given in Table 3.2.

Substituting the new aerodynamic resistance term into the Penman equation, and presenting the results as a water flux (kg of water per m^2 of area),

Table 3.2 Estimated values of aerodynamic and stomatal resistance for different vegetation types

Vegetation type	*Aerodynamic resistance (r_a) (s/m)*	*Canopy resistance (r_c) (s/m)*
Pasture	30	50
Forest	6.5	112
Scrub	6.5	160
Tussock	7.0	120

NB although the values of canopy resistance are presented as fixed they actually vary considerably throughout a day and season.
Source: from Andrew and Dymond (2007)

the evaporation estimation equation can be written as Equation 3.10.

$$PE = \frac{Q^* \Delta + \frac{\rho \cdot c_\rho \cdot \delta_e}{r_a}}{\lambda(\Delta + \gamma)} \tag{3.10}$$

where:

- Q^* = net radiation (W/m^2)
- Δ = rate of increase of the saturation vapour pressure with temperature (kPa/°C) (see Figure 3.6)
- ρ = density of air (kg/m^3)
- c_ρ = specific heat of air at constant pressure (≈ 1,005 J/kg)
- δ_e = vapour pressure deficit of the air (kPa)
- λ = latent heat of vaporisation of water (J/kg) (see Figure 3.4)
- γ = psychrometric constant (≈0.063 kPa/°C)
- r_a = aerodynamic resistance to transport of water vapour (s/m) given by Equation 3.11.

$$r_a = \frac{\left(In\left(\frac{z-d}{z_0} \right) \right)^2}{\kappa^2 u} \tag{3.11}$$

and

- κ = Von Karman constant (≈ 0.41)
- u = wind speed above canopy (m/s)
- z = height of anemometer (m)
- d = **zero plane displacement** (the height within a canopy at which wind speed drops to zero, often estimated at two-thirds of the canopy height) (m)
- z_o = roughness length (often estimated at one eighth of vegetation height) (m)

Although this formula looks complicated it is actually rather simple. It is possible to split the equation into two separate parts that conform to the understanding of evaporation already discussed. The available energy term is predominantly assessed through the net radiation (Q^*) term. Other terms in the equation relate to the ability of the atmosphere to absorb the water vapour (Δ, ρ, c_ρ, δ_e, λ, γ, this is referred to as the sensible heat transfer function) and the rate at which diffusion will absorb the water vapour into the atmosphere (κ, u, z_o, etc.).

Figure 3.6 shows the relationship between the saturated vapour pressure and temperature. The slope of this curve (Δ) is required in the Penman equation and its derivatives. This can be estimated from Equation 3.12 using average air temperature (T, °C):

$$\Delta = \frac{2053.058 \exp^{\frac{17.27T}{T+237.3}}}{(T+237.3)^2} \tag{3.12}$$

When using the Penman equation there are only four variables requiring measurement: net radiation, wind speed above the canopy, atmospheric humidity and temperature, which when combined will provide vapour pressure deficit (see Figures 3.7–3.9). Every other term in the equation is either a constant, a simple relationship from another variable or can be measured once. Of these four variables net radiation is the hardest to obtain from meteorological stations as net radiometers are not common. There are methods of estimating net

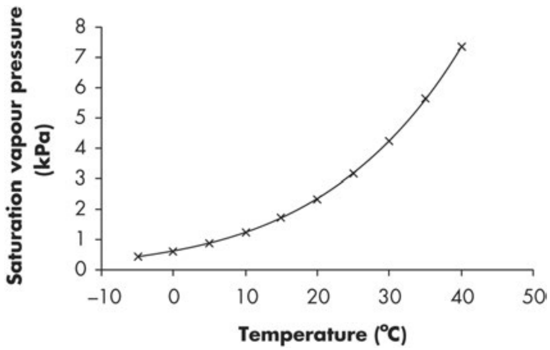

Figure 3.7 The relationship between temperature and saturation vapour pressure. This is needed to calculate the rate of increase of saturation vapour pressure with temperature (Δ). Equation 3.12 describes the form of this relationship.

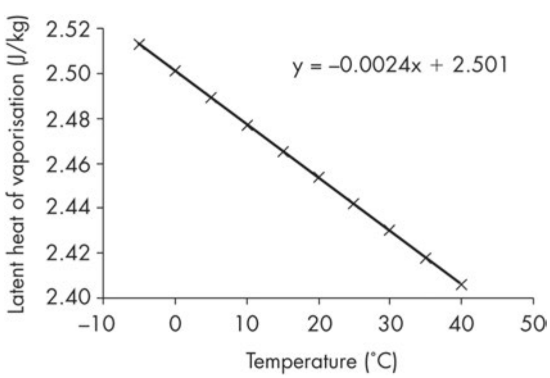

Figure 3.8 The relationship between temperature and latent heat of vaporisation.

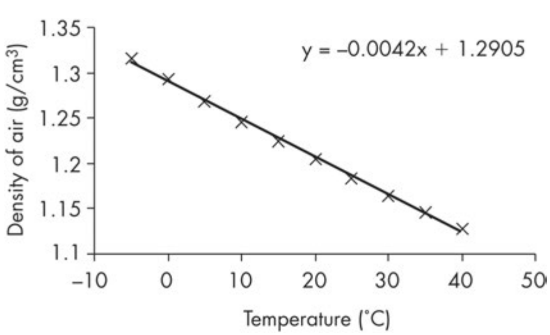

Figure 3.9 The relationship between air temperature and the density of air.

radiation from measurements of incoming solar radiation, surface **albedo** (or reflectivity) and day length (see Oke 1987).

The modified Penman equation provides estimates of potential evaporation at a surface for time intervals much less than the monthly value from Thornthwaite. This makes it extremely useful to hydrology and it is probably the most widely used method for estimating potential evaporation values.

However, there are problems with the Penman equation which make it less than perfect as an estimation technique. The assumption is made that the soil heat flux is unimportant in the evaporation energy budget. This is often the case but is an acknowledged simplification that may lead to some overall error, especially when the time step is less than one day. It is normal practice to use Penman estimates at the daily time step; however, in some modelling studies they are used at hourly time steps.

One major problem with the Penman equation relates to its applicability in a range of situations and in particular in the role of advection, as discussed on p. 51. This is where there are other energy sources available for evaporation that cannot be assessed from net radiation. Calder (1990) shows the results from different studies in the UK uplands where evaporation rates vastly exceed the estimates provided by the Penman equation. The cause of this discrepancy is the extra energy provided by cyclonic storms coming onto Britain from the Atlantic Ocean, something that is poorly accounted for in the Penman equation. The part of the Penman equation dealing with the ability of the atmosphere to absorb the water vapour (sensible heat transfer function) does account for some advection but not if it is a major energy source driving evaporation and it is highly sensitive to the aerodynamic resistance term. This does not render the Penman approach invalid; rather, in applying it the user must be sure that net radiation is the main source of energy available for evaporation or the aerodynamic resistance term is well understood.

Simplifications to Penman

There have been several attempts made to simplify the Penman equation for widespread use. Slatyer and McIlroy (1961) separated out the evaporation caused by sensible heat and advection from that caused by radiative energy. Priestly and Taylor (1972) derived a simplified Penman formula for use in the large-scale estimation of evaporation, in the order of 'several hundred kilometres' where it can be argued that large-scale advection is not important. Their formula for potential evaporation is shown in Equation 3.13.

$$PE = \alpha \frac{\left(Q^* - Q_G\right)\Delta}{\lambda(\Delta + \gamma)} \tag{3.13}$$

where Q_G is the soil heat flux term (often ignored by Penman but easily included if the measurements are available) and α is the Priestly–Taylor parameter, all other parameters being as defined earlier. The α term is an approximation of the sensible heat transfer function and was estimated by Priestly and Taylor (1972) to have a value of 1.26 for saturated land surfaces, oceans and lakes – that is to say, the sensible heat transfer accounts for 26 per cent of the evaporation over and above that from net radiation. This value of α has been shown to vary away from 1.26 (e.g. α = 1.21 in Clothier et al. 1982) but to generally hold true for large-scale areas without a water deficit. The Priestly–Taylor adaptation of the Penman equation is important for regional scale evapotranspiration assessment using satellite remote sensing (see Case Study in the Remote Sensing of Evaporation section).

Penman–Monteith

Monteith (1965) derived a further term for the Penman equation so that actual evaporation from a vegetated surface could be estimated. His work involved adding a canopy resistance term (r_c) into the Penman equation so that it takes the form of Equation 3.14.

$$E_t = \frac{Q^*\Delta + \rho c_\rho \delta_e / r_a}{\lambda\left(\Delta + \gamma\left(1 + \frac{r_c}{r_a}\right)\right)} \tag{3.14}$$

Looking at the Penman–Monteith equation you can see that if r_c equals zero then it reverts to the Penman equation (i.e. actual evaporation equals potential evaporation). If the canopy resistance is high the actual evaporation rate drops to less than potential. Canopy resistance represents the ability of a vegetation canopy to control the rate of transpiration. This is achieved through the opening and closing of stomata within a leaf, hence r_c is sometimes referred to as stomatal resistance. Various researchers have established canopy resistance values for different vegetation types (e.g. Szeicz et al. 1969), although they are known to vary seasonally and in some cases diurnally. Rowntree (1991) suggests that for grassland under non-limiting moisture conditions, the range of r_c should fall somewhere between 60 and 200 s/m. The large range is a reflection of canopy resistance being influenced by a plant's physiological response to variations in climatological conditions (see earlier discussion of stomatal control p. 54). Some values of canopy resistance for different vegetation types are given in Table 3.2.

Reference evaporation

The Penman–Monteith equation is probably the best evapotranspiration estimation method available, and is the method recommended by the Food and Agriculture Organisation (FAO). However for widespread use there is a need to have the stomatal resistance and aerodynamic resistance terms measured for a range of canopy covers at different stages of growth. To overcome this, the idea of reference evaporation has been introduced. This is the evaporation from a particular vegetation surface and the evaporation rate for another surface is related to this by means of crop coefficients. The Food and Agriculture Organisation (FAO) convened a group of experts who decided that the best surface for reference evaporation is close-cropped, well-watered grass. This is described in Allen et al. (1998) as a hypothetical reference crop with an assumed crop height of 0.12 m, a fixed canopy resistance of 70 s/m and an albedo of 0.23. Using these fixed values within the Penman–Monteith equation the reference evaporation (ET_o in mm/day) can be calculated from Equation 3.15.

$$ET_o = \frac{0.408\Delta\left(Q^* - Q_G\right) + \gamma \cdot \dfrac{900}{T+273} u \cdot \delta_e}{\Delta + \gamma\left(1 + 0.34_u\right)} \quad (3.15)$$

where:

Q^* is net radiation at the crop surface (MJ/m^2/day)
Q_G is soil heat flux density (MJ/m^2/day)
T is mean daily air temperature at 2 m height (°C)
u is wind speed at 2 m height (m/s)
δ_e is the saturation vapour pressure deficit (kPa)
Δ is the slope of the vapour pressure curve (kPa/°C)
γ is the psychrometric constant (kPa/°C)

The reference evapotranspiration provides a standard to which evapotranspiration at different periods of the year or in other regions can be compared and evapotranspiration of other crops can be related (Allen et al. 1998). Scotter and Heng (2003) have investigated the sensitivity of the different inputs to the reference evaporation equation in order to show what accuracy of measurement is required.

Table 3.3 outlines some crop coefficients as set out by FAO (Allen et al. 1998). At the simplest level the evapotranspiration for a particular crop can be estimated by multiplying the crop coefficient with the reference evapotranspiration although there are more complex procedures outlined in Allen et al. (1998) which account for growth throughout a season and climatic variability. Where the crop coefficient values shown in Table 3.3 are higher than 1.0 it is likely that the aerodynamic roughness of the canopy makes for higher evaporation rates than for short grass. Where the values are less than 1.0 then the plants are exerting stomatal control on the transpiration rate.

Simple estimation of E_t from *PE* and soil moisture

Where there is no stomatal control exerted by plants (e.g. in a pasture) the relationship between actual evaporation (E_t) and potential evaporation (PE) is by and large driven by the availability of water. Over a land surface the availability of water can be estimated from the soil moisture content

Table 3.3 Crop coefficients for calculating evapotranspiration from reference evapotranspiration

Crop type	*Crop coefficient (K_c)*	*Comment*
Beans and peas	1.05	Sometimes grown on stalks reaching 1.5 to 2 m in height. In such cases, increased K_c values need to be taken.
Cotton	1.15–1.20	
Wheat	1.15	
Maize (corn)	1.15	
Sugar cane	1.25	
Grapes	0.7	
Conifer forests	1.0	Conifers exhibit substantial stomatal control. The K_c can easily reduce below the values presented, which represent well-watered conditions for large forests.
Coffee	0.95	

(see Chapter 4). At a simple level it is possible to estimate the relationship between potential and actual evaporation using soil moisture content as a measured variable (see Figure 3.10). In Figure 3.10 a value of 1 on the y-axis corresponds to actual precipitation equalling potential evaporation (i.e. available water is not a limiting factor on the evaporation rate). The exact position where this occurs will be dependent on the type of soil and plants on the land surface, hence the lack of units shown on the x-axis and the two different curves drawn. This type of simple relationship has been effective in determining actual evaporation rates in a model of soil water budgeting (e.g. Davie et al. 2001) but cannot be relied on for accurate modelling studies. It provides a very crude estimate of actual evaporation from knowledge of soil moisture and potential evaporation.

The relationship between actual evaporation and soil moisture is not so simple where there is a vegetation type that exerts stomatal control on the evaporation rate (e.g. coniferous forest). In this case the amount of evaporation will be related to both soil moisture (available water) and the vapour pressure deficit (ability of the atmosphere to absorb water vapour). This is illustrated by Figure 3.11, a time series of soil moisture, transpiration and vapour pressure deficit for a stand of *Pinus radiata* in New Zealand. Transpiration was measured using sap-flow meters on a range of trees; soil moisture was

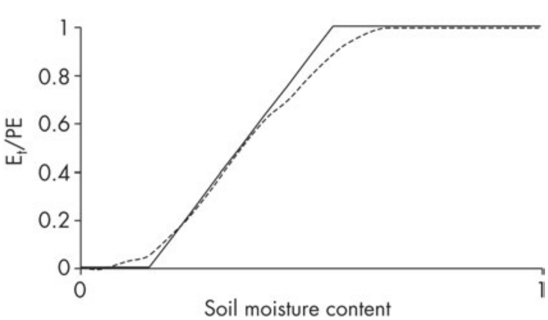

Figure 3.10 A hypothetical relationship between the measured soil moisture content and the ratio of actual evaporation to potential evaporation.

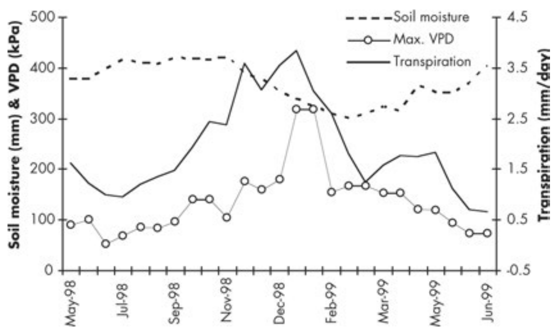

Figure 3.11 Time series of measured transpiration, measured soil moisture and estimated vapour pressure deficit for a forested site, near Nelson, New Zealand. NB as a Southern Hemisphere site the summer is from December until February.

Source: Data courtesy of Rick Jackson

measured with a neutron probe and vapour pressure deficit was estimated from a nearby meteorological station. At the start of the summer period (October to November 1998) the soil moisture level is high and the transpiration rate climbs rapidly to a peak. Once it has reached the peak, the transpiration rate plateaus, despite the maximum vapour pressure deficit continuing to climb. During this plateau in transpiration rate the forest is exerting some stomatal control so that the transpiration doesn't increase by as much as the vapour pressure deficit. From January 1999 (the height of the Southern Hemisphere summer) the transpiration rate drops markedly. Initially this matches a drop in the maximum vapour pressure deficit but the transpiration rate continues to drop below early summer rates (with similar VPD values). This is the time that the lack of soil moisture is starting to limit the tree transpiration. Figure 3.11 illustrates the complex relationship between evaporation from a vegetated surface, the soil moisture conditions and the atmospheric conditions.

Remote sensing of evaporation

Water vapour is a greenhouse gas and therefore it interferes with radiation (i.e. absorbs and reradiates) from the earth's surface. Because of this the amount of water vapour in the atmosphere can be estimated using satellite remote sensing, particularly using passive microwave sensors. The difficulty with using this information for hydrology is that it is at a very large scale (often continental) and is concerned with the whole atmosphere, not the near surface. In order to utilise satellites for estimation of evaporation, a combined modelling and remote sensing approach is required. Burke et al. (1997) describe a combined Soil–Vegetation–Atmosphere–Transfer (SVAT) model that is driven by remotely sensed data. This type of approach can be used to estimate evaporation rates over a large spatial area relatively easily. Mauser and Schädlich (1998) provide a review of evaporation modelling at different scales using remotely sensed data, while Zhang et al. (2016) provide a review of more recent developments. The Case Study below gives a more contemporary example of the use of remotely sensed data in estimating global evaporation.

Case study

SATELLITE REMOTE SENSING OF EVAPORATION

GLEAM (Global Land Evaporation Amsterdam Model) is a good example of trying to harness the maximum potential of satellite-derived data, using a variety of satellite sensor products. First developed in 2011 (Miralles et al. 2011), it has been successively revised and at the time of writing a third version (GLEAM v3) was available (Martens et al. 2017) (see below for a summary of the datasets that have emerged from each version). GLEAM is a set of algorithms that individually estimate each of the components of evaporation from land. These are open water evaporation and sublimation (from snow and ice), transpiration and bare earth evaporation, and interception loss.

GLEAM uses the Priestly–Taylor equation (refer back to the Simplifications to Penman section of this chapter) to calculate potential evaporation for water bodies and areas covered by ice and snow because most of the required data can be directly observed by satellites. The data required for this are observations of surface net radiation and near-surface air temperature, both of which can be measured by satellites such as NASA's Terra and Aqua. The Priestly–Taylor equation is also used for calculating evaporation from areas of bare soil and land covered by vegetation; in the case of the latter, this is for two categories of vegetation (tall canopy and short canopy). As you will know from the earlier part of this chapter, for vegetated areas potential evaporation needs to be converted to actual evaporation, which in turn depends on vegetation and the soil moisture available to these plants. In GLEAM, this is done by integrating two datasets, both of which originate from the European Space Agency's Soil Moisture Ocean Salinity (SMOS) mission. The first is the microwave-based Vegetation Optical Depth and the second is an estimate of soil moisture. Flux in soil moisture is modelled separately in another set of algorithms in GLEAM (the soil module), which has a multi-layer representation of the soil profile and is driven by precipitation derived from several other satellites. To minimise errors in this modelling process, the SMOS moisture observations are also assimilated into the model. Another set of algorithms (stress module) contains semi-empirical relationships relating stress and soil moisture condition in the root zone for vegetation characterised by the Vegetation Optical Depth. These determine the stress factor to convert potential evaporation into actual evaporation. The last of the evaporation components mentioned at the start of this case study, interception loss, assesses evaporation directly off the surface of vegetation after it has been intercepted during rainfall (Chapter 4 will cover the process of interception in more detail). In GLEAM, interception evaporation is estimated by the Gash model and driven by observed rainfall. Although this component has been mentioned last, it is modelled first because it determines the effective rainfall influencing the soil moisture.

What this case study illustrates is that although evaporation can be modelled using remotely sensed data, doing so requires integration of a number of satellite data products and process based hydrological models. New generations of satellite are likely to improve current efforts. At the time of writing, three GLEAM v3.1 datasets were available (GLEAM website: www.gleam.eu), at a scale of 0.25°:

1. **GLEAM_v3.1a**: a global dataset spanning the 37-year period 1980–2016. The dataset is based on reanalysis of net radiation and air temperature, satellite and gauged-based precipitation, VOD, soil moisture, and snow water equivalent.
2. **GLEAM_v3.1b**: a 50°N–50°S dataset spanning the 13-year period 2003–2015 and driven exclusively by satellite data.
3. **GLEAM_v3.1c**: a 50°N–50°S dataset spanning the 5-year period 2011–2015 and driven exclusively by satellite data, including VOD and soil moisture from SMOS.

Further information on GLEAM and access to these datasets on a non-commercial basis is available at the GLEAM website (www.gleam.eu).

Mass balance estimation

In the same manner that evaporation pans and lysimeters estimate evaporation rates, evaporation at the large scale (catchment or lake) can be estimated through the water balance equation. This is a relatively crude method, but it can be extremely effective over a large spatial and/or long temporal scale. The method requires accurate measurement of precipitation and runoff for a catchment or lake. In the case of a lake, change in storage can be estimated through lake-level recording and knowledge of the surface area. For a catchment it is often reasonable to assume that change in storage is negligible over a long time period (e.g. 1 year) and therefore the evaporation is precipitation minus runoff.

EVAPORATION IN THE CONTEXT OF WATER QUANTITY AND QUALITY

Evaporation, as the only loss away from the surface in the water balance equation, plays a large part in determining available water quantity. The loss of water from soil through direct evaporation and transpiration has a direct impact on the amount of water reaching a stream during high rainfall (see Chapter 7) and also the amount of water able to infiltrate through soil and into groundwater (see Chapters 5 and 6). The impact of evaporation on water quantity is not as great as for precipitation but it does have a significant part to play in the quantity and timing of water flowing down a river.

The influence of evaporation on water quality is mostly through the impurities left behind after water has evaporated. This may lead to a concentration of impurities in the water remaining behind (e.g. the Dead Sea between Israel and Jordan) or a build-up of salts in soils (salination). This is discussed in more detail in Chapter 10.

ESSAY QUESTIONS

1. **Give a detailed account of the factors influencing evaporation rate above a forest canopy.**
2. **Compare and contrast the use of evaporation pans and lysimeters for measuring evaporation.**

3 Outline the major evaporation estimation techniques and compare their effectiveness for your local environment.

4 Describe the factors that restrict actual evaporation (evapotranspiration) from equalling potential evaporation in a humid-temperate climate.

FURTHER READING

Allen, R.G., Pereira L.S., Raes, D. and Smith, D. (1998) Crop evapotranspiration – Guidelines for computing crop water requirements. *FAO Irrigation and drainage paper 56* (available at www.fao.org/documents).

Brutsaert, W. (1982) *Evaporation into the atmosphere: Theory, history, and applications*. Kluwer, Dordrecht.
A detailed overview of the evaporation process.

Calder, I.R. (1990) *Evaporation in the uplands*. J. Wiley & Sons, Chichester.
Although concerned primarily with upland evaporation, it covers the issues of estimation well.

Cheng, M. (2003) *Forest hydrology: An introduction to water and forests*. CRC Press, Boca Raton, F.L.
An overview of forest hydrological processes, including evaporation and interception loss.

4

INTERCEPTION AND SURFACE STORAGE

LEARNING OBJECTIVES

When you have finished reading this chapter you should have:

- An understanding of the importance of surface storage, particularly in relation to storage by snow and ice.
- A knowledge of how storage in snow and ice is measured and estimated.
- An understanding of the processes of precipitation interception by vegetation.
- An understanding of the techniques for estimating interception loss and measuring other processes involved in interception.
- An understanding of depression storage as an initial abstraction.

The water balance equation, explained in Chapter 1, contains a storage term (S). Within the global hydrological cycle there are several areas where water can be considered to be stored, most notably soil moisture, groundwater, snow and ice and, to a lesser extent, lakes and reservoirs. It is tempting to see stored water as static, but in reality there is considerable movement involved. The use of a storage term is explained in Figure 4.1 where it can be seen that there is an inflow, an outflow and a movement of water between the two. The inflow and outflow do not have to be equal over a time period; if not, then there has been a *change in storage* (ΔS). The critical point is that at all times there is some water stored, even if it is not the same water throughout a measurement period.

This definition of stored water is not perfect as it could include rivers as stored water in addition to groundwater, etc. The distinction is often made based on flow rates (i.e. how quickly the water moves while in storage). There is no critical limit to say when a deep, slow river becomes a lake, and likewise there is no definition of how slow the flow has to be before becoming stored water. It relies on an intuitive judgement that slow flow rates constitute stored water.

The importance of stored water is highlighted by the fact that it is by far the largest amount

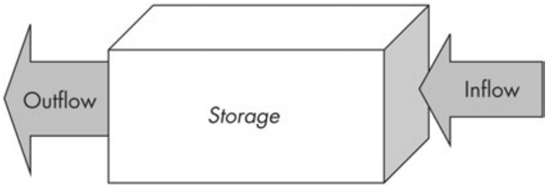

Figure 4.1 Illustration of the storage term used in the water balance equation.

of fresh water in or around planet earth (see Table 1.2, p. 8). The majority of this is as surface storage – either in snow and ice (particularly the polar ice caps) or groundwater (sub-surface storage). For many parts of the world, groundwater is a major source of drinking water, so knowledge of amounts and replenishment rates is important for water resource management (see Chapter 5). By definition, stored water is slow moving so it is particularly prone to contamination by pollutants. The three 'Ds' of water pollution control (dilution, dispersion and degradation; see Chapter 10) all occur at slow rates in stored water, making pollution management a particular problem. When this is combined with the use of these waters for potable supply, an understanding of the hydrological processes occurring in stored water is very important.

In this chapter we will focus on surface storage by snow and ice, the other two major sub-surface stores in the soil zone and in groundwater will be discussed in the following two chapters. This chapter will also consider the process of interception and interception storage and loss, as well as depression storage, and we will start with these processes.

INTERCEPTION

Interception refers to the process whereby the fall of precipitation to the ground is 'intercepted' or interrupted by being caught by vegetation. One of three things can happen to precipitation caught in this way. Firstly, it can be evaporated back into the atmosphere, thereby contributing to **interception loss**. Secondly, it can be held in the canopy until it is dislodged by wind or drips off leaves, eventually falling to the ground, in which case it is called **throughfall**. Finally, it can run down the stems of leaves and branches towards the trunk, and ultimately into the ground at the base of the trunk, and this is called **stemflow**. This is illustrated in Figure 4.2, and can happen at a forest canopy, at the level of shrubs or grass, and even on a litter layer. Figure 4.3 represents this as a systems diagram, showing the fluxes and storages involved in these processes.

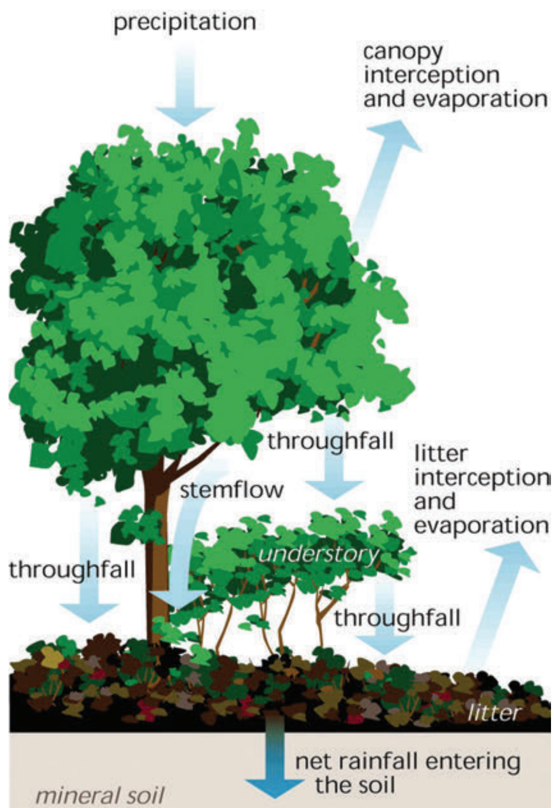

Figure 4.2 Processes and concepts in interception.

Source: Federal Interagency Stream Restoration Working Group (FISRWG) (1998)

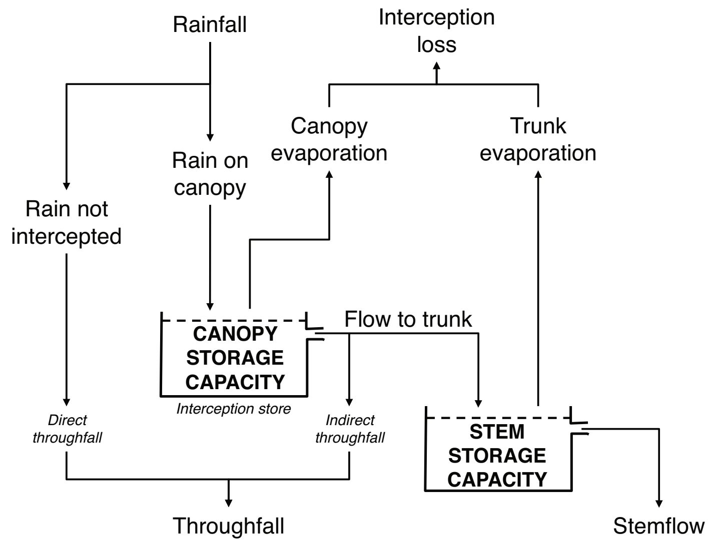

Figure 4.3 A systems diagram of the processes of interception.
Adapted from Ladson (2008)

Throughfall

This is the water that falls to the ground either directly, through gaps in the canopy (*direct throughfall*), or indirectly, having dripped off leaves, stems or branches (*indirect throughfall*). The amount of *direct throughfall* is controlled by the canopy coverage for an area, a measure of which is the leaf area index (LAI). LAI is the ratio of leaf area to ground surface area and consequently has a value greater than one when there is more than one layer of leaf above the ground. When the LAI is less than one you would expect some direct throughfall to occur. When you shelter under a tree during a rainstorm you are trying to avoid the rainfall and direct throughfall. The greater the surface area of leaves above you, the more likely it is that you will avoid getting wet from direct throughfall.

The amount of *indirect throughfall* is also controlled by the LAI, in addition to the **canopy storage capacity** and the rainfall characteristics. Canopy storage capacity is the volume of water that can be held by the canopy before water starts dripping as indirect throughfall, referred to by some as the **interception store**. The canopy storage capacity is controlled by the size of trees, plus the area and water-holding capacity of individual leaves. The latter would be influenced by factors such as the shape, orientation, roughness, and flexibility of leaves, including features such as the presence of fine hairs. Seasonality is important too, as canopy storage would vary between trees that shed their leaves in winter (deciduous) and evergreen trees.

Although vegetation characteristics are important in understanding interception, climatic factors are equally important including rainfall rate, wind and evaporative demand. For example, rainfall characteristics are an important control on indirect throughfall as they dictate how quickly the canopy storage capacity is filled. Referring back to the earlier analogy of standing under a tree during a rainstorm, experience should tell you that intensive rainfall quickly turns into indirect throughfall (i.e. you get wet!), whereas light showers frequently do not reach the ground surface at all.

In reality, canopy storage capacity is a rather nebulous concept. Canopy characteristics are constantly changing and it is rare for water on a canopy to fill up completely before creating indirect throughfall. This means that indirect throughfall occurs before the amount of rainfall equals the canopy storage capacity, making it difficult to gauge exactly what the storage capacity is.

Stemflow

Stemflow is the rainfall that is intercepted by stems and branches and flows down the tree trunk into the soil. Although measurements of stemflow show that it is a small part of the hydrological cycle (normally 2–10 per cent of above canopy rainfall (Lee 1980)) it can have a much more significant role. Durocher (1990) found that trees with smoother bark such as beech (*Fagus*) had higher rates of stemflow as the smoothness of bark tends to enhance drainage towards stemflow.

Stemflow acts like a funnel, collecting water from a large area of canopy but delivering it to the soil in a much smaller area: the surface of the trunk at the base of a tree. This is most obvious when considering large canopy leafy trees, but it still applies for other structures (e.g. conifers) where the area of stemflow entry into the soil is far smaller than the canopy catchment area for rainfall. At the base of a tree it is possible for the water to rapidly enter the soil through flow along roots and other macropores surrounding the root structure. This can act as a rapid conduit of water sending a significant pulse into the **soil water**.

Interception loss

While water sits on the canopy, prior to indirect throughfall or stemflow, it is available for evaporation. When this evaporation occurs, this is known as *interception loss*. You will recall this was mentioned briefly in the previous chapter, and also in the Case Study on the use of remote sensing for the estimation of evaporation.

Interception gain

In some circumstances it is possible that there is an interception gain from vegetation. In the Bull Run catchment, Oregon, USA it has been shown that the water yield after timber harvesting was significantly less than prior to the trees being logged (Harr 1982; Ingwersen 1985). This is counter to the majority of catchment studies reported by Bosch and Hewlett (1982) which show an increase in water yield as forests are logged. The reason for the loss of water with the corresponding loss of trees in Oregon is to do with the particular circumstances of the catchment. Fog from the cold North Pacific, with no accompanying rain, is a common feature and the trees intercept fog particles, creating 'fog drip' which is a significant part of the water balance. Fog droplets are extremely small, and Ingwersen (1985) has suggested that the sharp ends of needles on pine trees act as condensation nuclei, promoting the growth of larger droplets that fall to the ground (see an example of fogdrip from tussock leaves in Figure 3.5). When the trees are removed there are no condensation nuclei (or far fewer) on the resultant vegetation so the water remains in the atmosphere and is 'lost' in terms of water yield. Equally important is the influence of vegetation roughness. The turbulent mixing of air as wind passes over a rough canopy promotes rapid deposition of condensing water (directly converse to interception loss; see Chapter 3). The overall result of this is that the removal of trees in the Oregon case leads to less water in the river; this runs counter to the evidence provided in the Maimai Case Study in Chapter 7.

It is the role of interception loss (wet leaf evaporation) that makes afforested areas greater users of water than pasture land (see Case Study on p. 74). This is because the transpiration rates are similar between pasture and forest, but the interception loss is far greater from a forested area. There are two influences on the amount of interception loss from a particular site: canopy structure and meteorology. Canopy structural factors include the storage capacity, the drainage characteristics of the canopy and the aerodynamic roughness of the canopy. The morphology of leaf and bark on a tree are important factors in controlling how quickly water drains towards the soil. If leaves are pointed upwards then there tends to be a rapid drainage of water towards the stem. Sometimes this appears as an evolutionary strategy by a plant in order to harvest as much water as possible (e.g. rhubarb and gunnera plants). Large broadleaved plants, such as oak (*Quercus*), tend to hold water well on their leaves, while needled plants can hold less per leaf (although they normally have more leaves). Seasonal changes make a large difference within deciduous forests, with far greater interception losses when the trees have leaves than without. Table 4.1 illustrates the influence of plant morphology through the variation in interception found in different forest types and ages.

The largest influence that a canopy has in the evaporation process is through the aerodynamic roughness of the top of the canopy. This means that as air passes over the canopy it creates a turbulent flow that is very effective at moving evaporated water away from the surface. The reason that forests have such high interception losses is because they have a lot of intercepting surfaces *and* they have a high aerodynamic roughness leading to high rates of diffusion of the evaporated water away from the leaf (Figure 4.4).

Meteorological factors affecting the amount of interception loss are the rainfall characteristics. The rate at which rainfall occurs (intensity) and storm duration are critical in controlling the interception loss. The longer water stays on the canopy the greater the amount of interception loss. Also

Table 4.1 Interception measurements in differing forest types and ages

Tree type	*Age*	*Interception (mm)*	*% annual precipitation*
Deciduous hardwoods	100	254	12
Pinus strobus (White pine)	10	305	15
Pinus strobus	35	381	19
Pinus strobus	60	533.4	26

Source: From Hewlett and Nutter (1969)

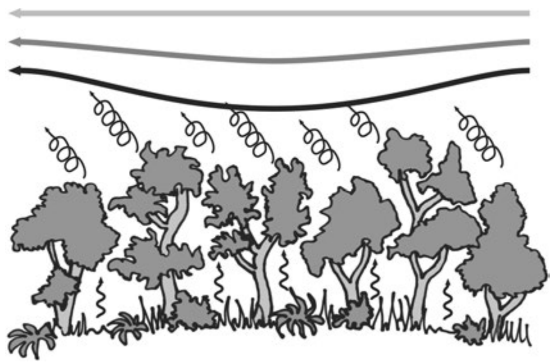

Figure 4.4 Factors influencing the high rates of interception loss from a forest canopy. The capacity of the leaves to intercept rainfall and the efficient mixing of water vapour with the drier air above leads to high evaporative losses (interception loss).

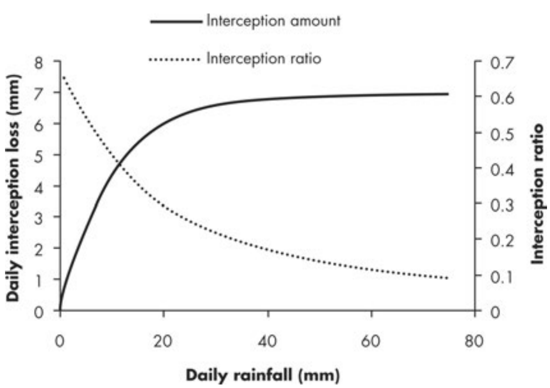

Figure 4.5 Empirical model of daily interception loss and the interception ratio for increasing daily rainfall. An interception ratio of 1.0 means all rainfall becomes interception loss.

Source: Redrawn from Calder (1999)

important will be the frequency of rainfall. Does the canopy have time to dry out between rain events? If so, then the interception amount is likely to be higher. This is demonstrated in Figure 4.5 where the percentage of interception loss (interception ratio – broken line) is higher for small daily rainfall totals and the actual interception amount (solid line) reaches a maximum value of around 7 mm even in the largest of daily rainfalls.

The amount of interception loss from an area is climate dependent. Calder (1990) used an amalgamation of different UK forest interception studies to show that there is a higher interception ratio (the interception loss divided by above-canopy rainfall) in drier than in wetter climates. The interception ratio ranges from 0.45 at 500 mm annual rainfall, to 0.27 at 2,700 mm annual rainfall. It is important to note that these interception ratio figures have considerable inter-annual variability.

Case study

FORESTS AND RAINFALL VS EVAPORATION

If you stand watching a forest during a warm summer shower it is common enough to see what appear to be clouds forming above the trees. For many years it was believed that somehow trees attract rainfall and that cloud-forming was evidence of this phenomenon. As described by Pereira (1989) 'The worldwide evidence that hills and mountains usually have more rainfall and more natural forests than do adjacent lowlands has historically led to confusion of cause and effect'. This idea was taken further so that it became common practice to have forestry as a major land use in catchments that were being used to collect water for potable supply. In fact, the cloud formation that is visible above a forest is a result of evaporation occurring from water sitting on the vegetation (intercepted rainfall). This 'wet leaf evaporation' can be perceived as a loss to the hydrologist as it does not reach the soil surface and contribute to possible streamflow. Throughout the latter half of the twentieth century there was considerable debate on how important wet leaf evaporation is.

One of the first pieces of field research to promote the idea of canopy interception being important was undertaken at Stocks Reservoir, Lancashire, UK. Law (1956) studied the water balance of an area covered with conifers (Sitka spruce) and compared this to a similar area covered with grassland. The water balance was evaluated for areas isolated by impermeable barriers with evaporation left as the residual (i.e. rainfall and runoff were measured and soil moisture assumed constant by looking at annual totals). Law found that the evaporation from the forested area was far greater than that for the pasture and he speculated that this was caused by wet leaf evaporation – in particular that the wet leaf evaporation was far greater from the forested area as there was a greater storage capacity for the intercepted water. Furthermore, Law went on to calculate the amount of water 'lost' to reservoirs through wet leaf vegetation and suggested a compensation payment from the forestry owners to water suppliers.

Conventional hydrological theory at the time suggested that wet leaf evaporation was not an important part of the hydrological cycle because it compensated for the reduction in transpiration that occurred at the same time (e.g. Leyton and Carlisle 1959; Penman 1963). In essence it was believed that the evapotranspiration rate stayed constant whether the canopy was wet or dry.

Following the work of Law, considerable research effort was directed towards discovering whether the wet leaf/dry leaf explanation was responsible for discrepancies in the water balance between grassland and forest catchments. Rutter (1967) and Stewart (1977) found that wet leaf evaporation in forests may be up to three or four times that from dry leaf. In contrast to this, other work has shown that on grassland, wet leaf evaporation is approximately equal to dry leaf (McMillan and Burgy 1960; McIlroy and Angus 1964). In addition, transpiration rates for pasture have been found to be similar to that of forested areas. When all this evidence is added up it confirms Law's work that forested areas 'lose' more rainfall through evaporation of intercepted water than grassland areas.

However, there is still a question over whether the increased wet leaf evaporation may lead to a higher regional rainfall – a form of water recycling. Bands et al. (1987) write that: 'Forests are associated with high rainfall, cool slopes or moist areas. There is some evidence that, on a continental scale, forests may form part of a hydrological feedback loop with evaporation contributing to further rainfall'. Most researchers conclude that in general there is little, if any, evidence that forests can increase rainfall. However, Calder (1999: 24, 26) concludes, 'Although the effects of forests on rainfall are likely to be relatively small, they cannot be totally dismissed from a water resources perspective. . . . Further research is required to determine the magnitude of the effect, particularly at the regional scale.'

Interception measurement

The most common method of assessing the amount of canopy interception loss is to measure the precipitation above and below a canopy and assume that the difference is from interception. Stated in this way it sounds a relatively simple task but in reality it is fraught with difficulty and error. Durocher (1990) provides a good example of the instrumentation necessary to measure canopy interception, in this case for a deciduous woodland plot.

Above-canopy precipitation

To measure above-canopy precipitation, a rain gauge may be placed on a tower above the canopy. The usual rain gauge errors apply here, but especially the exposure to the wind. As described earlier in this chapter, the top of a forest canopy tends to be rough and is very good for allowing turbulent transfer of evaporated water. The turbulent air is not so good for measuring rainfall! An additional problem for any long-term study is that the canopy is not static; the tower needs to be raised every year so that it remains above the growing canopy.

One way around the tower problem is to place a rain gauge in a nearby clearing and assume that what falls there is the same amount as directly above the canopy nearby. This is often perfectly reasonable to assume, particularly for long-term totals, but care must be taken to ensure the clearing is large enough to avoid obstruction from nearby trees (see Figure 2.10).

Throughfall

Throughfall is the hardest part of the forest hydrological cycle to measure. This is because a forest canopy is normally variable in density and therefore throughfall is spatially heterogeneous. One common method is to place numerous rain gauges on the forest floor in a random manner. If you are interested in a long-term study then it is reasonable to keep the throughfall gauges in fixed positions. However, if the study is investigating individual storm events then it is considered best practice to move the gauges to new random positions between storm events. In this way the throughfall catch should not be influenced by gauge position. To derive an average throughfall figure it is necessary to come up with a spatial average in the same manner as for areal rainfall estimates.

To overcome the difficulty of a small sampling area (rain gauge) measuring something notoriously variable (throughfall), some investigators have used either troughs or plastic sheeting. Troughs collect over a greater area and have proved to be very effective (see Figure 4.6). Plastic sheeting is the ultimate way of collecting throughfall over a large area, but has several inherent difficulties. The first is purely logistical in that it is difficult to install and maintain, particularly to make sure there are no rips. The

Figure 4.6 Throughfall troughs sitting beneath a pine tree canopy. This collects rain falling through the canopy over the area of the trough. It is sloping so that water flows to a collection point.

second is that by having an impervious layer above the ground there is no, or very little, water entering the soil. This might not be a problem for a short-term study but is over the longer term, especially if the investigator is interested in the total water budget. It may also place the trees under stress through lack of water, thus leading to an altered canopy.

Stemflow

The normal method of measuring stemflow is to place collars around a tree trunk that capture all the water flowing down the trunk. On trees with smooth bark this may be relatively simple but is very difficult on rough bark such as found on many conifers. It is important that the collars are sealed to the tree so that no water can flow underneath and that they are large enough to hold all the water flowing down the trunk. The collars should be sloped to one side so that the water can be collected or measured in a tipping-bucket rain gauge. Maintenance of the collars is very important as they easily clog up or become appropriate resting places for forest fauna such as slugs!

Interception estimation

As is evident from the preceding section, measurement of interception is very challenging. Firstly, there are the challenges in measuring above canopy rainfall and secondly there are the problems in measuring what eventually reaches the ground – so that you can then infer the difference as being the part of the interception storage that was lost via evaporation. As you can imagine, each vegetation type will have its own particular challenges in measurement, and in many types this will vary seasonally. For this reason, empirical studies reported in the literature will focus on a particular type of vegetation. Because of their importance and also because of the significance of interception in the local hydrology of forested areas, most work in this area has been on forests.

Early estimation approaches were based on simple regression type methods which linked interception loss to falling precipitation, while Horton (1919) proposed interception loss would be comprised of the intercepted rain evaporated during the storm plus that remaining after the storm, and which would later evaporate (Robinson and Ward 2017). Empirical models that link rainfall to interception loss based on regression relationships of measured data sets have been developed for many different types of vegetation canopy (see Zinke (1967) and Massman (1983) for examples and reviews of these types of model). Some of these models used logarithmic or exponential terms in the equations but they all rely on having regression coefficients based on the vegetation type and climatic regime.

A more detailed modelling approach is the Rutter model (Rutter et al. 1971, 1975) which calculates an hourly water balance within a forest stand. The water balance is calculated, taking into account the rate of throughfall, stemflow, interception loss through evaporation and canopy storage. In order to use the model, a detailed knowledge of the canopy characteristics is required. In particular the canopy storage and drainage rates from throughfall are required to be known; the best method for deriving these is through empirical measurement. The Rutter model treats the canopy as a single large leaf, although it has been adapted to provide a three-dimensional canopy (e.g. Davie and Durocher 1997) that can then be altered to allow for changes and growth in the canopy. Because the Rutter model operates on a short time-step, requires much data and is complex to use, it is generally only used in a research context (Robinson and Ward 2017).

A simplified approach which retains much of the correct process representation of the Rutter model was proposed by Gash (1979). This method distinguishes between the vegetation and meteorological factors controlling interception and calculates interception for each storm event. Subsequent empirical testing found that both the Gash and Rutter methods overestimated for sparser canopies in temperate areas leading to the reformulated Gash model (Gash et al. 1995) which is now the most widely implemented approach (Muzylo et al. 2009). The latter authors provide a useful review of interception modelling.

Although remote sensing techniques can provide useful information that can be incorporated into canopy interception models, these methods cannot provide the detailed difference between above- and below-canopy rainfall. The value of remote sensing is that satellites can give good information on the type of vegetation and its degree of cover. However, particular care needs to be taken over the term 'leaf area index' (LAI) when reading remote sensing literature. Analysis of remotely sensed images can provide a good indication of the percentage vegetation cover for an area, but this is not necessarily the same as leaf area index – although it is sometimes referred to as such. Leaf area index is the surface area of leaf cover above a defined area divided by the surface area defined. As there are frequently layers of vegetation above the ground, the leaf area index frequently has a value higher than one. The percentage vegetation cover cannot exceed one (as a unitary percentage) as it does not consider the third dimension (height).

Nevertheless as both satellite sensor technology and analytical methods develop, we can expect to see progress in the estimation of interception by remote sensing methods. For example, Miralles et al. (2011) produced the first global estimate of cumulative rainfall interception over a 5-year period (2003–2007), giving some insight into both the spatial variability and magnitude of forest interception at a global scale. This method adapted the Gash model and used satellite derived precipitation, land cover and lightning frequency (to identify where higher intensity rainfall was occurring) in deriving estimates. More recently, Cui and Jia (2014) have developed a remote sensing Gash model (RS-Gash) where remote sensing observations of Vegetation Area Index (VAI) and Fractional Vegetation Cover (FVC) were used to assess interception loss from heterogeneous forest at a regional scale in the Heihe river basin, China. Comparison with ground measurements showed reasonable accuracy in estimation, providing further evidence that the Gash model can be used successfully at a regional scale. Interception loss is of course a component of the evaporation budget, and as noted in the Case Study on remote sensing of evaporation (p. 66), the GLEAM model for estimating global evaporation (Martens et al. 2017) uses the Gash model implemented in using the same approach presented in Miralles et al. (2011).

DEPRESSION STORAGE

While the preceding section has focused on interception by vegetation, precipitation can also be intercepted and stored by other surfaces. Consider for example, the numerous flat impervious surfaces in an urban environment. If precipitation here is not drained effectively it will be stored and subject to evaporation. This is not limited to the urban environment either; there are many other places where precipitation may be held on the surface. In agricultural landscapes, for example, water will pond in small depressions. Although some might infiltrate the soil, where soils are compacted or surface crusting has occurred this will be limited and stored water will be subject to evaporation. Knowing something about these storages (and potential losses) is important as they influence the proportion of precipitation that eventually becomes runoff.

SNOW AND ICE STORAGE

The parts of the earth's surface where water exists in solid form (sea, lake and river ice, glaciers, snow cover, ice caps, ice sheets and frozen ground) is known as the *cryosphere*. Snow and ice are hugely important stores of water for many countries in the world, particularly at high latitudes or where there are large mountain ranges. The gradual release of water from snow and ice (glaciers), either during spring and summer or on reaching a lower elevation, makes a significant impact on the hydrology of many river systems.

In the same manner that rainfall may be intercepted by a canopy, so can snow. The difference between the two is in the mass of water held and the duration of storage (Lundberg and Halldin 2001). The amount of intercepted snow is frequently much higher than for rainwater and it is held for much longer. This may be available for evaporation through sublimation (moving directly from a solid to a gas) or release later in snow melt. Hedstrom and Pomeroy (1998) point out that the mass of snow held by interception is controlled by the tree branching structure, leaf area and tree species. In countries such as Canada and Russia there are extensive forests in regions dominated by winter snowfall. Some studies have shown as much as 20–50 per cent of gross precipitation being intercepted and evaporated (Lundberg and Halldin 2001). These figures indicate that a consideration of snowfall interception is critical in these regions. Lundberg and Halldin (2001) provide a review of measuring snow interception and modelling techniques.

The timing of snow and ice melt is critical in many river systems, but especially so in rivers that flow north towards the Arctic Circle. In this case the melting of snow and ice may occur in the headwaters of a river before it has cleared further downstream (at higher latitudes). This may lead to a build-up of water behind the snow and ice further downstream – a snow and ice dam (see Figure 4.7). Beltaos (2000) estimates that the cost of damage caused by ice-jams in Canada is around $60 million dollars per annum. The Case Study below gives an example of the flow regime that may result from this type of snow melt event.

Figure 4.7 Susquehanna river ice jam and flood which destroyed the Catawissa Bridge in Pennsylvania, USA on 9 March 1904.

Source: Photo copyright of Columbia County Historical and Genealogical Society

Case study

THE MACKENZIE RIVER: DEMONSTRATING THE INFLUENCE OF SNOW AND ICE ON RIVER FLOWS

The Mackenzie river is at the end of one of the great river systems of the world. In North America the Missouri/Mississippi system drains south, the Great Lakes/Hudson system drains eastwards, and the Mackenzie river system drains northwards with its mouth in the Beaufort Sea (part of the Arctic Ocean). The Mackenzie river itself has a length of 1,800 km from its source: the Great Slave Lake. The Peace and Athabasca rivers which flow into the Great Slave Lake, and are therefore part of the total Mackenzie drainage basin, begin in the Rocky Mountains 1,000 km to the southwest. The total drainage basin is approximately 1,841,000 km^2, making it the twelfth largest in the world, and the river length is 4,250 km (see Figure 4.8).

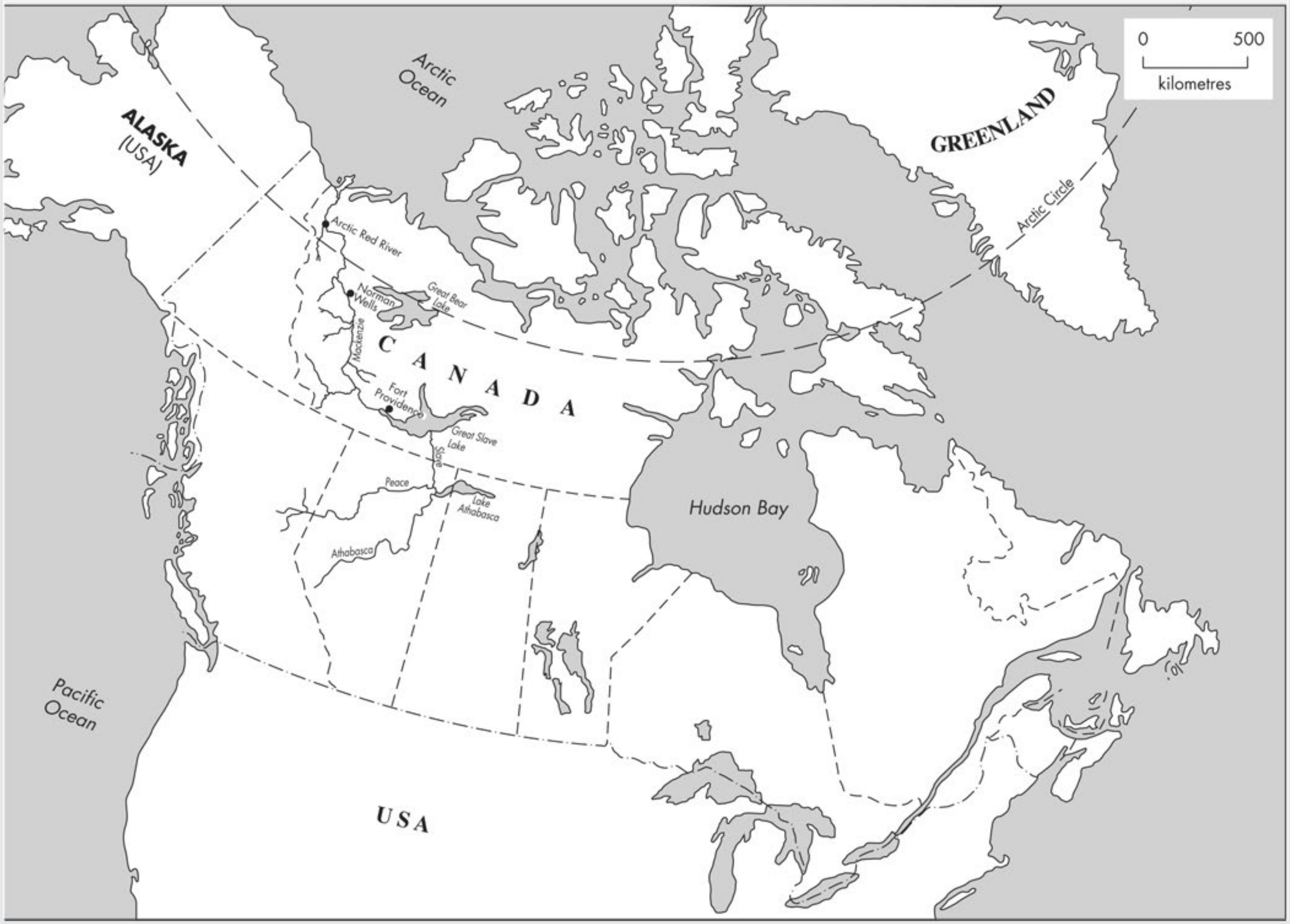

Figure 4.8 Location of the Mackenzie river in Canada.

The high latitude of the Mackenzie river makes for a large component of snow and ice melt within the annual hydrograph. This can be seen in Figure 4.9, taken at the junction of the Mackenzie and Arctic Red rivers. This gauging station is well within the Arctic circle and towards the delta of the Mackenzie river. The monthly discharge values increase dramatically from April to June when the main melt occurs and then gradually decrease to reach a minimum value during the winter months. The highest average streamflow occurs in June despite the highest precipitation occurring one month later in July. Overall there is very little variation in precipitation but a huge variation on riverflow. This is an excellent example of the storage capability of snow and ice within a river catchment. Any water falling during the winter months is trapped in a solid form (snow or ice) and may be released only during the warmer summer months. The amount of precipitation falling during the summer months (mostly rainfall) is dwarfed by the amount of water released in the melt during May and June.

The most remarkable feature of a river system such as the Mackenzie is that the melt starts occurring in the upper reaches, sending a pulse of water down the river, before ice on the lower reaches has melted properly. This is not unique to the Mackenzie river, all the great rivers draining northwards in Europe, Asia and North America exhibit the same tendencies. If we look in detail at an individual year (Figure 4.10), you can see the difference in daily hydrographs for stations moving down the river (i.e. northwards). Table 4.2 summarises the information on latitude and

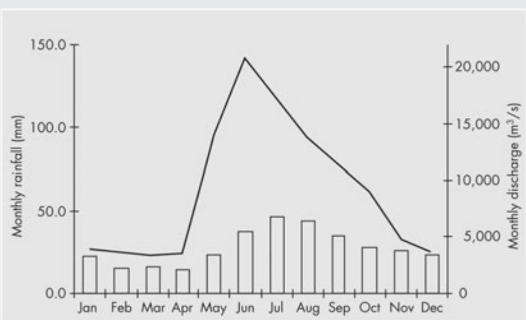

Figure 4.9 Average monthly river flow (1972–1998; line) for the Mackenzie river at the Arctic Red River gauging station (latitude 67° 27′ 30″ N) and average precipitation (1950–1994) for the Mackenzie river basin (bars).

Source: Data courtesy of Environment Canada

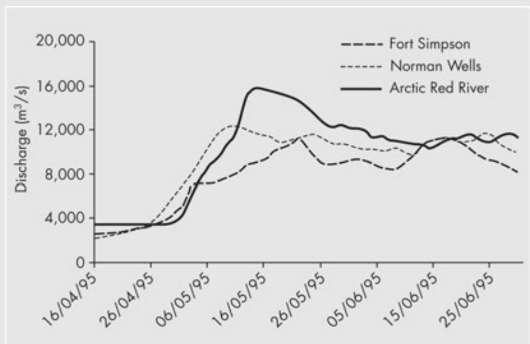

Figure 4.10 Daily river flow at three locations on the Mackenzie river from mid-April through to the end of June 1995.

Source: Data courtesy of Environment Canada

Table 4.2 Summary of latitude and hydrological characteristics for three gauging stations on the Mackenzie river

Mackenzie river gauging station	*Latitude (north)*	*Date of last ice on river (1995)*	*Date of peak discharge (1995)*
Fort Simpson	61° 52′ 7″	14 May	21 May
Norman Wells	65° 16′ 26″	18 May	10 May
Arctic Red River	67° 27′ 30″	31 May	14 May

Source: Data courtesy of Environment Canada flow characteristics for the Mackenzie. It is not a simple story to decipher (as is often the case in hydrology), but you can clearly see that the rise in discharge at the Arctic Red River station starts later than the Norman Wells station further to the south. The rise is caused by melt, but predominantly from upstream. It is also clear that for both Norman Wells and the Arctic Red River stations the highest discharge value of the year is occurring while the river is still covered in ice. This creates huge problems for the drainage of the area as the water may build up behind an ice dam. Certainly, the water flowing under the ice will be moving much quicker than the ice and water mix at the surface. Figure 4.11 demonstrates the way that water builds up behind an ice dam, particularly where there is a constriction on either side of the river channel.

Figure 4.11 Ice dam forming on the Mackenzie river, Canada.

Source: Photograph courtesy of Dr Faye Hicks, University of Alberta

Measuring snow depth

The simplest method of measuring snow depth is the use of a core sampler. This takes a core of snow, recording its depth. The snow sample can then be melted to derive the water equivalent depth, the measurement of most importance in hydrology. The major drawbacks in using a core sampler to derive snow depth are that it is a non-continuous reading (similar to daily rainfall measurement) and the position of coring may be critical (because of snow drifting).

A second method of measuring snow depth is to use a *snow pillow*. This is a sealed plastic pillow that is normally filled with some form of antifreeze and connected to a pressure transducer (Figure 4.12). When left out over a winter period the weight of snow on top of the pillow is recorded as an increasing pressure, which can be recalculated into a mass of snow. It is important that the snow pillow does not create an obstacle to drifting snow in its own right. To overcome this, and to lessen the impact of freezing on the pillow liquid, it is often buried under a shallow layer of soil or laid flat on the ground. When connected to a continuous logging device a snow pillow provides the best record of snow depth (and water equivalent mass).

Estimating snow cover

The main method of estimating snow cover is using satellite remote sensing. Techniques exist that give a reasonably good method of detecting the areal extent of snow cover, but it is much harder to translate this into a volume of snow (i.e. by knowing depth) or water. Optical and thermal infrared data can be used to estimate snow cover but they rely heavily on the reflective ability of snow; unfortunately, other surfaces such as clouds may also exhibit these properties (Fitzharris and McAlevey 1999). Microwave data offer a far better method of detecting snow cover. Passive microwaves detected by a satellite can be interpreted to give snow cover because any water (or snow) covering the surface absorbs some of the microwaves emitted by the earth surface. The greater the amount of snow the weaker is the microwave signal received by the

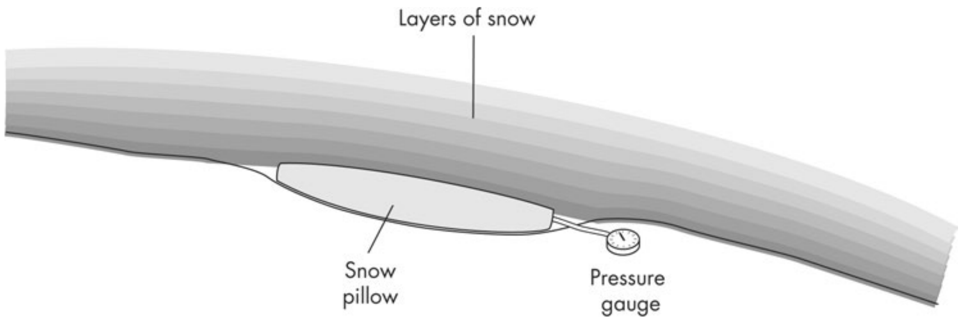

Figure 4.12 Snow pillow for measuring weight of snow above a point. The snow weight is recorded as a pressure exerted on the pillow.

satellite. Ranzi et al. (1999) have used AVHRR imagery to monitor the snowpack in an area of northern Italy and some of Switzerland and compared this to other measurement techniques. Earlier passive microwave satellite sensors (e.g. AVHRR) were at an extremely coarse spatial resolution that is really only applicable at the large catchment scale. Active microwave sensing offers more hope but its usage for detecting snow cover is still being developed (see Case Study). Dietz et al. (2012) give a detailed review of remote sensing of snow, including spatial and temporal resolution of sensors and processing algorithms. Frei (2012) provides a useful review of global satellite derived snow products.

Snow melt

Of critical importance to hydrology is the timing of snow melt, as this is when the stored water is becoming available water. There are numerous models that have been developed to try and estimate the amount of snow melt that will occur. Ferguson (1999) gives an earlier summary of snow melt modelling work. The models can be loosely divided into those that rely on air temperature and those that rely on the amount of radiation at a surface. The former frequently use a degree days approach, the difference between mean daily temperature and a melting threshold temperature. Although it would seem sensible to treat zero as the melting threshold temperature this is not always the case; snow will melt with the air temperature less than zero because of energy available through the soil (soil heat flux) and solar radiation. The degree day snow melt approach calibrates the amount of snow that might be expected given a certain value of degree day. Although this is useful for hydrological studies it is often difficult to calibrate the model without detailed snow melt data. More recently, Clark et al. (2011) review approaches to representing snow inputs in hydrological and land-surface models, which are strongly dependent on the spatial scale at which processes are being represented.

Case study

USING REMOTELY SENSED DATA TO MONITOR STORAGE IN SNOW AND ICE

The use of satellite imagery and passive remote sensing (RS) to monitor surface water, snow and ice is well established, starting in 1970s with the inclusion of non-visible spectral bands in the early LandSat Multispectral Scanner. Characteristic reflectances of snow, ice and water were identified in laboratory studies and it was hoped this information would inform the image interpretation process for identifying such features. Unfortunately, the initial optimism about such approaches remained largely unrealised in practice, as the complexity and variability of the real world introduced elements of both doubt and error into the products of image interpretation. However, this does not mean that modern RS techniques have no value – the science has progressed significantly in the 30 years since the launch of LandSat-1 in 1978 – but rather that a cautious, and meticulously documented approach to RS products and their use is needed.

Identifying ice and snow

Estimates of volume of water in surface ice and snow are important in streamflow and water storage forecasting, which in turn, affects things like hydroelectric power generation potential and irrigation programmes, and many other activities that have an economic impact.

One of the problems with RS approaches to identifying ice and snow, is separating these features spectrally from cloud. However, there are characteristics that do distinguish them from clouds, particularly, higher absorption in the middle-infrared. What is problematic, though, is that the reflectance of both snow and ice changes in different conditions, for example, the high reflectance of fresh snow (think bright white) decreases as the snow ages, and characteristics such as density, particle size, the amount of liquid water, and impurity levels change. Ice reflectance changes too, depending on its roughness, age, thickness, and the presence of air bubbles. Notwithstanding these difficulties, many snow and ice observation projects have been implemented based on passive remote sensing products.

The fact that snow, ice and cloud are also likely to occur at the same time in a given location, meant that in the early days of RS assessments ice and snow were unusual, even after the inclusion of near-infrared channels on sensors. It was only once sensors in the middle-infrared were deployed that separation between these features became possible – because clouds have considerably higher reflectance than ice and snow in this region of the electromagnetic spectrum. Imagery from LandSat, Spot and Sentinel could be used for assessments of snow and ice, within the constraints outlined above.

An alternative approach uses passive microwave remote sensing and allows assessments of snow, area, depth, and/or snow water equivalent albeit at a coarse spatial resolution (from 12.5 km to 150 km). These are derived from a variety of satellites and sensors and are available for download as spatial time series at a variety of spatial and temporal resolutions (see section below).

A number of RS missions have focused primarily on monitoring the polar ice caps. Between 2003 and 2010, NASA's ICESat (Ice, Cloud and land Elevation Satellite) mission (using a satellite-based laser altimeter) collected multi-year elevation data which allowed the determination of ice sheet balance, as well as topography and global vegetation data. A follow-up mission is underway with the Launch of ICESat-2 on 15 September 2018.

The ESA initiated a similar mission in 1998, called CryoSat, but due to a problem during the launch in 2005, it failed to achieve orbit and was lost. Within 5 months, a replacement mission was approved and CryoSat-2 launched in 2010. CryoSat-2 was the first satellite to carry a synthetic aperture radar (SAR) on board. It provides data about the polar ice caps, and tracks changes in thickness with an accuracy of 1.3 cm. These datasets are proving critical in tracking the loss of polar ice in response to climate change.

Identifying water

The characteristic reflectance and absorbency characteristics of water allow its presence on the planet's surface to be assessed easily using RS approaches.

Water on the surface of the planet can have a range of characteristics, for example, it could be carrying a heavy sediment load, or perhaps have a vigorous algal bloom, and it can, of course, vary in depth. Visually, the first example might be browny-red if we were to look at it, while the second might be bright green or red. So, looking from space these water features might look, spectrally, like bare earth, or perhaps fields of young crops, if we were to view them using only the visible parts of the electromagnetic spectrum. Similarly, very shallow water might look more like the depression containing it, than like water at all.

The key characteristic of water for RS purposes is its distinctive absorption of energy in the near-infrared and beyond; consequently, the relatively high spatial resolution, frequency of observation, and ease of data access of a range of passive visible/near-infrared sensors has meant that sensors such as MODIS, the LandSat family of sensors, and more recently Sentinel-2's MSI are widely used in surface water focused projects. They allow easy assessment of the surface area of water features – but subject to the constraints of cloud cover associated with all optical imagery.

Active microwave remote sensing is also used to identify water on the earth's surface, and has the advantage of not being impacted by cloud cover. The mapping of water surfaces with radar is based on the different backscatter regimes of land and water surfaces. Smooth water surfaces have low returns, whereas the surrounding land is much rougher, leading to high backscatter. This response, combined with the fact that such active RS technologies are capable of penetrating the atmosphere under virtually all conditions, make it very useful for rapid assessment of areas affected by floods.

Knowing where water is, is important; knowing how deep it is, is another key element of interest. This may be related to a flooding situation as mentioned above, but could also, through the analysis of a time-series of images, allow the monitoring of water levels in lakes and impoundments of ecological and economic importance. In 2004, the European Space Agency launched a project called Monitoring River and Lake Levels from Space. It aimed to provide accurate and easy-to-use river and lake level data to the scientific community derived from the radar-based ERS and Envisat satellite altimeters. A description of this initiative is available from www.esa.int/esapub/bulletin/bullet117/chapter5_bul117.pdf.

Data sources and classification methods

Water features can be extracted either through a threshold approach in a single band, or through the application of an index involving a combination of bands such as the Normalized Difference Vegetation Index (NDVI) or the Normalized Difference Water Index (NDWI) – exploiting the multispectral elements of the imagery. Similarly, the Normalized Difference Snow Index (NDSI) can be used to identify areas of snow cover.

The National Snow and Ice Data Center (https://nsidc.org/) maintains and permits access to a record of passive microwave imagery from 1978 to present, as well as access to a range of related research reports and data related to glaciers, ice sheets, sea ice and snow.

Information about the NASAs ICESat and ICESat-2 missions, and associated data can be accessed from: https://icesat.gsfc.nasa.gov/index.php.

LandSat data can be downloaded from the USGS's EarthExplorer website (https://earthexplorer.usgs.gov/), while Sentinel data is accessible from the Copernicus Open Access Hub (https://scihub.copernicus.eu/).

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ESSAY QUESTIONS

1 Describe a field experiment (including equipment) required to measure the water balance beneath a forest canopy.

2 Select two contrasting vegetation types and explain how interception processes may differ.

3 Discuss the challenges in quantifying storage in snow and ice.

4 Explain the relevance of scale in determining surface storage.

FURTHER READING

Chang, M. (2012) *Forest hydrology: An introduction to water and forests* (3rd edition). CRC Press, Boca Raton, F.L.
A more modern text than Lee (1980) which gives a good overview of forest hydrology.

DeWalle, D. (2011) *Forest hydrology*. IAHS Benchmark Papers in Hydrology Series. IAHS Press, Wallingford.
Forming part of the IAHS 'benchmark paper' series, this is a compendium of historic papers and associated commentary reflecting the evolution of understanding in the field.

Lee, R. (1980) *Forest hydrology*. Columbia University Press, New York.
A classic text giving an excellent overview of forest hydrological processes.

Robinson, M. and Ward, R. (2017) Chapter 3: Interception. In: *Hydrology: Principles and processes*. IWA publishing, London.
An excellent overview of interception, providing more detail on interception for different vegetation types.

5

GROUNDWATER

LEARNING OBJECTIVES

When you have finished reading this chapter you should have:

- An understanding of the importance of groundwater in the hydrological cycle.
- An understanding of the factors that determine (i) the direction and (ii) the rate of groundwater movement.
- An understanding of the key concepts in groundwater hydrology and the properties of aquifers.
- A knowledge of the techniques for measuring groundwater.
- An understanding of the importance of groundwater as a store and as a water supply.

In the previous chapters, we have focused on processes above the earth's surface. In this and the following chapter, we will consider water beneath the surface. Water occurs as *soil water* in the soil zone (also called the **vadose zone** or **zone of aeration**), and as groundwater below the **water table** (or **phreatic surface**) (the area below the water table is called the **phreatic zone**) (Figure 5.1). Although it might seem more logical to consider soil water first, there are good reasons why it is easier to understand water below the water table. This relates to the fact that below the water table *saturated* conditions exist, in contrast to the *unsaturated* conditions in the soil zone. Water movement under unsaturated conditions is more complicated than at saturation, so we will deal with groundwater in this chapter and then build on that understanding in the following chapter, where we will discuss soil water. The study of water in the saturated zone is a specialised discipline undertaken by *hydrogeologists* or *geohydrologists* (these terms are usually used interchangeably, although a hydrogeologist would normally have a stronger grounding in geology and vice versa). This chapter is intended as a very brief introduction to groundwater, for further reading see Younger (2007) as an excellent introductory text and Hiscock and Bense (2014), Fitts (2013), Schwartz and Zhang (2003), and Todd and Mays (2005), for more advanced texts.

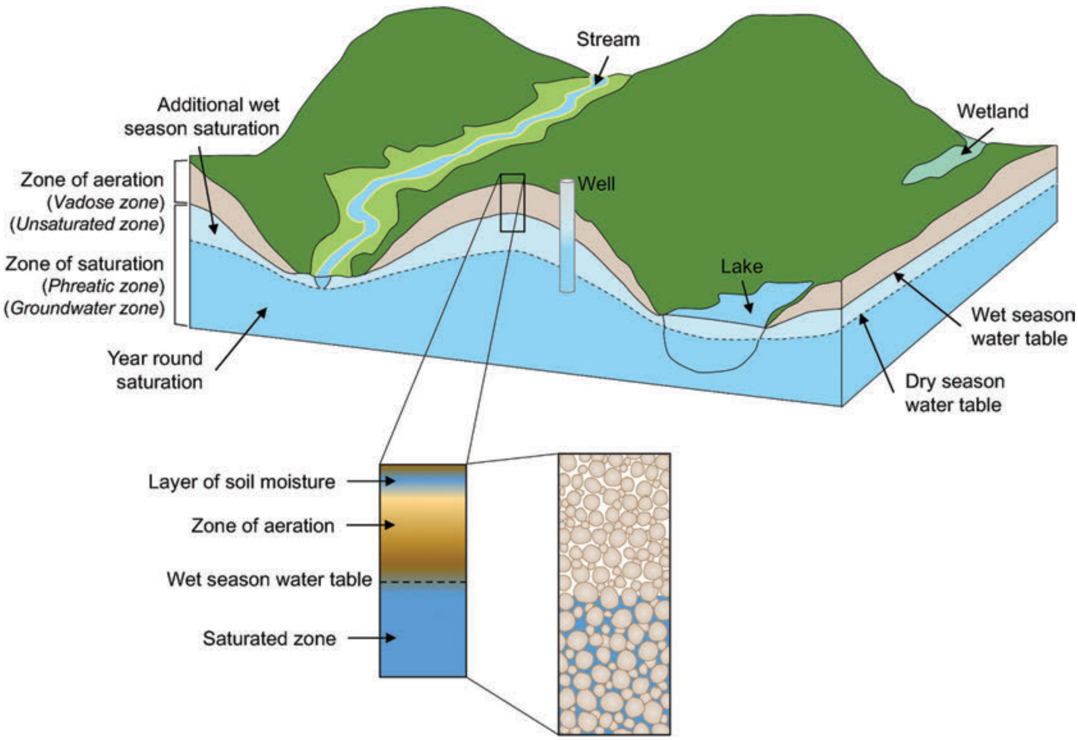


Figure 5.1 Water stored beneath the earth's surface. Rainfall passes through the unsaturated vadose zone to become groundwater – this process is known as groundwater *recharge*. The broken line represents the water table, but this is almost always a transition from saturated to unsaturated conditions, rather than an abrupt change. The water table generally mirrors the topography of the land surface, although in a much more muted way. In certain types of geology this zone of transition (also called a *capillary fringe* – we'll see why later), can be as deep as 1 m.

Source: Adapted from Skinner et al. (2004)

WATER STORAGE BELOW THE EARTH'S SURFACE

Soil forms a relatively thin covering on the earth's surface, typically less than 1–2 m deep, below which rock of various types is found. Often this occurs as a transition from soil, to weathered and fractured rock, and then to unweathered bedrock or **regolith**. So where is water stored below the surface? To understand this, we need to think about the nature of the materials forming soil and rock. Both soil and rock are composed of solid particles – if we lift a handful of soil we can clearly see this, but between these particles we can also see spaces. These spaces can, of course, be filled with water or air. This can be less obvious to the naked eye in rock, but is nevertheless the case. The implication of this is that when thinking about materials below the earth we need to think in relation to three components (i) *solids*, (ii) *pore spaces* (voids or open spaces between particles) and then (iii) whether these are filled with *air* or *water*. The simple answer to the question above of 'where is water stored?', is of course, in these pore spaces or *interstices*. Under saturated conditions *all* the available spaces are filled by water, whereas in the vadose zone these might be filled by just air, or air and some water, and sometimes only water. It is important to note that although we refer to the soil zone as 'unsaturated', this does not mean that it is never saturated; saturation can, and does, occur. The name therefore refers to the general condition. As we shall see, it is also not quite as straightforward as this – for example, it is not just about visible spaces, but also the microscopic spaces, including those within the apparently solid structure of the individual particles. The significant point to remember at this stage, however, is that the relationship between solid particles and spaces (or voids) is key in determining how much water can be stored in a soil or rock. If we are to understand groundwater, being able to quantify this relationship (and related processes) is obviously important. This is what we will try to do in the next section.

Porosity

A fundamental attribute of both soil and rock is its **porosity**. This is defined simply as the proportion of a soil or rock that consists of open spaces, and must therefore logically be the volume of the pores divided by the total volume of the rock or soil, often represented using the symbol 'n'. (Figure 5.2, Equation 5.1).

$$n = \frac{V_p}{V_t} \tag{5.1}$$

Where n is the porosity of a sample (dimensionless, or e.g. cm^3/cm^3), V_p is the volume of pores in the sample (e.g. cm^3) and V_t is the total volume of the sample (e.g. cm^3).

Porosity is usually expressed as a decimal fraction (between 0 and 1), but it can also be expressed as a percentage. Table 5.1 gives an indication of the typical range in porosity of different types of rock and soil. What is evident from this table is that it is largely unconsolidated deposits and sedimentary rocks that have high porosities. Metamorphic and igneous rocks have lower porosities – unless fracturing and weathering has subsequently taken place.

In civil engineering and soil mechanics, use is also made of a strongly related concept, that of the *void ratio* (usually symbolised as ε). Whereas porosity is the ratio of the volume of voids to the total volume, the voids ratio is the volume of the voids (empty or full of water) to the volume of just the solids (V_s) (Equation 5.2). The reason for this difference is related to the engineer's greater interest in the structural properties of materials.

$$\varepsilon = \frac{V_p}{V_s} \tag{5.2}$$

Another important difference between saturated and unsaturated conditions relates to pressure. Pore pressure arises from all the pores being full of water; this means that below the water table, pore water pressure is greater than atmospheric pressure. Above the water table, where pores contain a mixture of water and air, pore pressure is generally less than atmospheric pressure.

In unconsolidated materials (e.g. soils or alluvial floodplain deposits) the pore spaces are clearly evident, whereas they may be less obvious in hard rock. How rock was formed, and indeed what has happened to it subsequently, has an important bearing on porosity. Similarly, and also all things being equal, where the rock is in relation to others also has a bearing on porosity. For example, you might expect the same sandstone buried under layers of other rock to have less porosity, simply because the pressure above is likely to have compressed the particles and pore spaces of the lower strata of rock.

We can therefore distinguish between **primary porosity** and **secondary porosity**. Primary porosity is the porosity that exists at the time of formation of the rock. For example, alluvial sands in a flood plain might have a high primary porosity. As they are buried, compressed and over geological time become consolidated into sandstone, they might lose some porosity and acquire secondary porosity. Other types of rock might be changed by the percolation of mineral rich water which precipitates out in the voids, reducing the available space and therefore porosity. As indicated above, geological processes, particularly metamorphic processes,

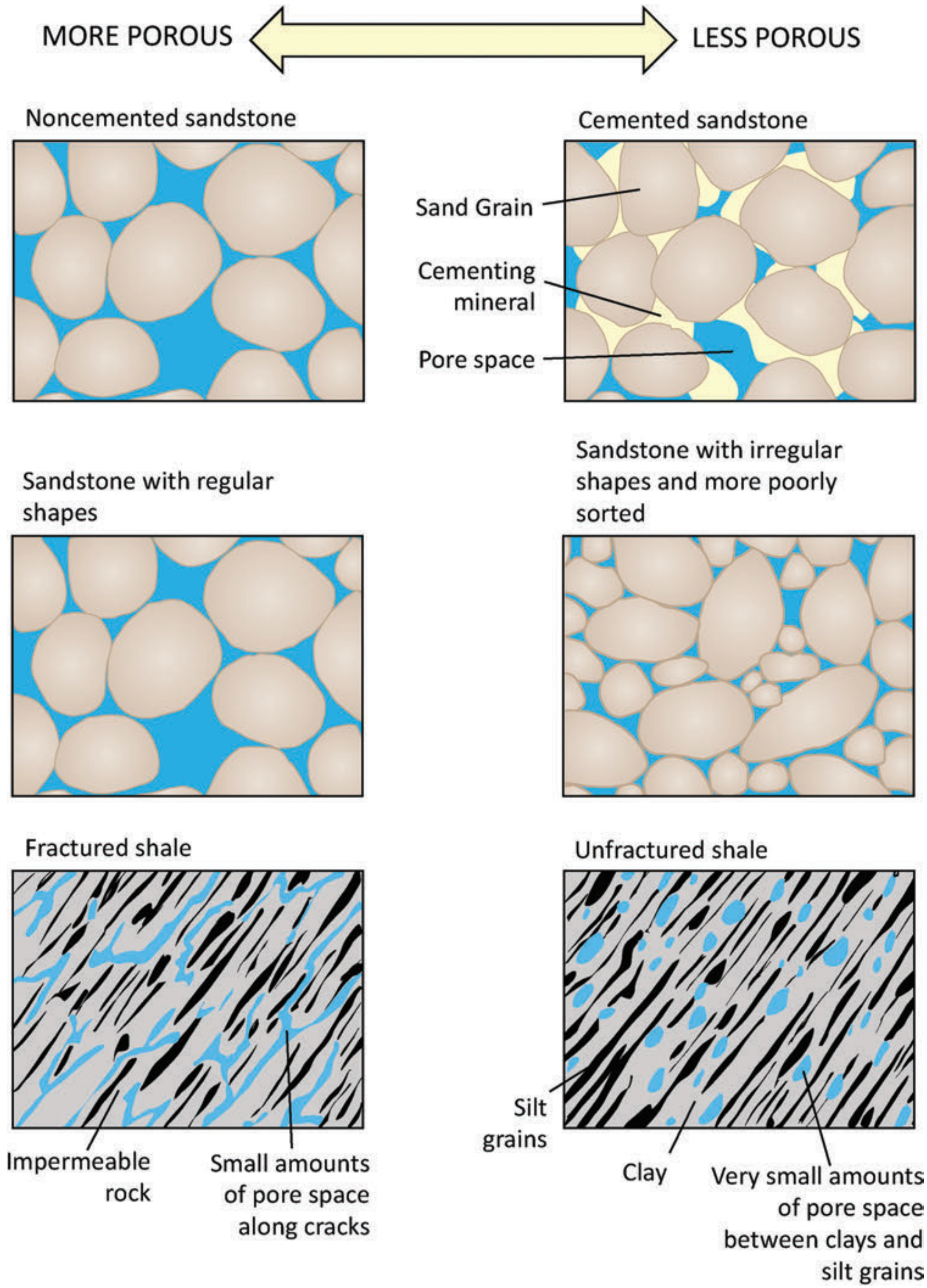


Figure 5.2 The concept of porosity, defined as the proportion of the total volume of a body of soil or rock that is made up of pore spaces. As illustrated, various factors determine the porosity of a material.

Table 5.1 Typical values of porosity (*n*) for different types of rock and soil

	Material	*Comment*	*Porosity (n) (%)*
Unconsolidated	Soils	Porosity is generally high in unconsolidated materials, depending on the degree of sorting and particle size distribution. These sediments are often associated with unconfined aquifers.	30–65
	Clay		35–60
	Silt		35–60
	Sand		30–50
	Sand – fine		26–53
	Sand – coarse		31–46
	Gravel		25–40
	Fluvial deposits (alluvium)		5–35
Sedimentary	Sandstone	Porosity is variable in sedimentary rocks. It can be very low in unfractured shales but moderate to high in siltstones, sandstones and some types of limestone. These types of rock can be aquifers or aquitards depending on their properties.	5–30
	Siltstone		20–40
	Karst limestone		5–50
	Chalk/oolitic limestone		5–30
	Dolomitic limestone		0–50
	Mudstone		35–45
	Unfractured shale		0–10
	Fractured shale		5–50
Igneous/metamorphic	Dense crystalline rocks	Porosity is generally very low or low in igneous and metamorphic rocks unless it has been fractured or weathered. These types of rock tend to be aquicludes or aquifuges.	0–5
	Crystalline rocks		0–10
	Basalt		3–35
	Slate		1–5
	Granite		0–10
	Fractured crystalline rocks		0–10
	Fractured marble		<2
	Weathered granite		35–55
	Weathered gabbro		42–45

Source: Adapted from Ward and Robinson (2000); Schwartz and Zhang (2003); Hiscock and Bense (2014)

can significantly change the characteristic of parent materials, rendering them harder and denser due to the pressures and temperatures they are subjected to. However, 'hard' igneous and metamorphic rocks are also subject to weathering, and in more geologically active zones, fracturing and faulting. These processes break down rocks at various scales, generating new materials with different porosities.

Another qualifier of our general understanding of porosity, is the important concept of **effective porosity**. The way to think about this is to imagine a rock where the water is included as completely separate little pockets – perhaps trapped there when the rock formed. Now think of a rock where these pockets are connected by smaller pathways of pores. In the first case, although we would say that the rock was able to store water, it wouldn't be much use to us as how would we get it out? In the second case the connected pore spaces would allow the flow of water, so this would be a much more useful situation as we could extract the water. We can distinguish between these two situations by using the idea of *total porosity*, as distinct from *effective porosity*. Whereas the former refers to the total number of voids, the latter refers only to the proportion of spaces that are *connected*. So total porosity says something about water storage in rock, effective porosity tells us about how *permeable* the rock is, or its ability to allow water to flow through. Fractured rock, for example, has very high effective porosity and can be an important pathway for groundwater flow.

These qualifiers of porosity are less relevant in the soil zone, but it is nevertheless important to note that porosity within a soil will vary depending on geological parent material, location in the soil profile and the history of soil forming processes at that location. We will focus on soils in the next chapter, so from here on we are focusing on the saturated zone only.

WATER STORED AS GROUNDWATER

What should be clear from the fundamental principles above is that how much water can be stored in rock or unconsolidated sediments depends very much on the type of rock. A body of rock or unconsolidated sediment that can store water is called an **aquifer**. Historically this really meant storage and transmission in sufficient quantities that could be abstracted for use, but nowadays extraction is not a requirement for it to be defined as such. Aquifers range in geology from unconsolidated gravels such as the Ogallala aquifer in the USA (see Chapter 11) to distinct geological formations (e.g. chalk underlying London and much of south-east England). As can be seen from Table 5.1, these are mostly sedimentary rocks and unconsolidated materials. In contrast, some rocks (e.g. unweathered igneous and metamorphic rocks) have very low porosities and do not store water. These strata (layers) or bodies of rock are called **aquifuges** (a relatively impermeable material that does not contain or transmit water) (Todd and Mays 2005). An **aquiclude** is a saturated, but relatively impermeable, material that does not yield much water. Saturated materials of intermediate water holding capacity or permeability (i.e. not as much as an aquifer, but more than an aquifuge and aquiclude) are known as **aquitards**, in other words they might *retard* the flow of water but they might not *exclude* it. These terms are applied in relation to both the storage and permeability characteristics of rock. However, these are 'relative' and not 'absolute' definitions – whether a body of rock is an aquifer or not depends on what it is next to (Ward and Robinson 2000). For example, a mudstone layer between two layers of sandstone would be an aquitard with aquifers on either side. However, the same mudstone layer adjacent to an igneous body could be an aquifer, as it would contain considerably more water than the igneous rock.

Types of aquifers

There are broadly two forms of aquifer type: **unconfined** and **confined**. An *unconfined* aquifer has no boundary above it and therefore the water table is free to rise and fall dependent on the amount of water contained in the aquifer (see Figure 5.3). The lower boundary of the aquifer may be impervious but it is the upper boundary, or water table, that is unconfined and may intersect the surface. A particular type of unconfined aquifer, is a **perched water table** or **perched aquifer** (see Figure 5.3) where an impermeable layer prevents the percolation of water down to the regional water table. Perched water tables may be temporary features reflecting variable hydraulic conductivities within the soil and rock, or they can be permanent features reflecting the overall geology. Quite often perched aquifers give rise to hillside seeps or small wetlands.

A *confined aquifer* has a flow boundary (aquitard) above and below it that constricts the flow of water into a confined area (see Figure 5.3). Geological formations are the most common form of confined aquifers, and as they often occur as layers, the flow of water is restricted in the vertical dimension but not in the horizontal. Water within a confined aquifer is normally under pressure and if intersected by a borehole will rise up higher than the constricted boundary. If the water reaches the earth's surface it is referred to as an **artesian well**. A little later we will consider what might determine the height to which the water might rise. Related to this point, note the line marked 'potentiometric surface' in Figure 5.3 – we'll come back to this later.

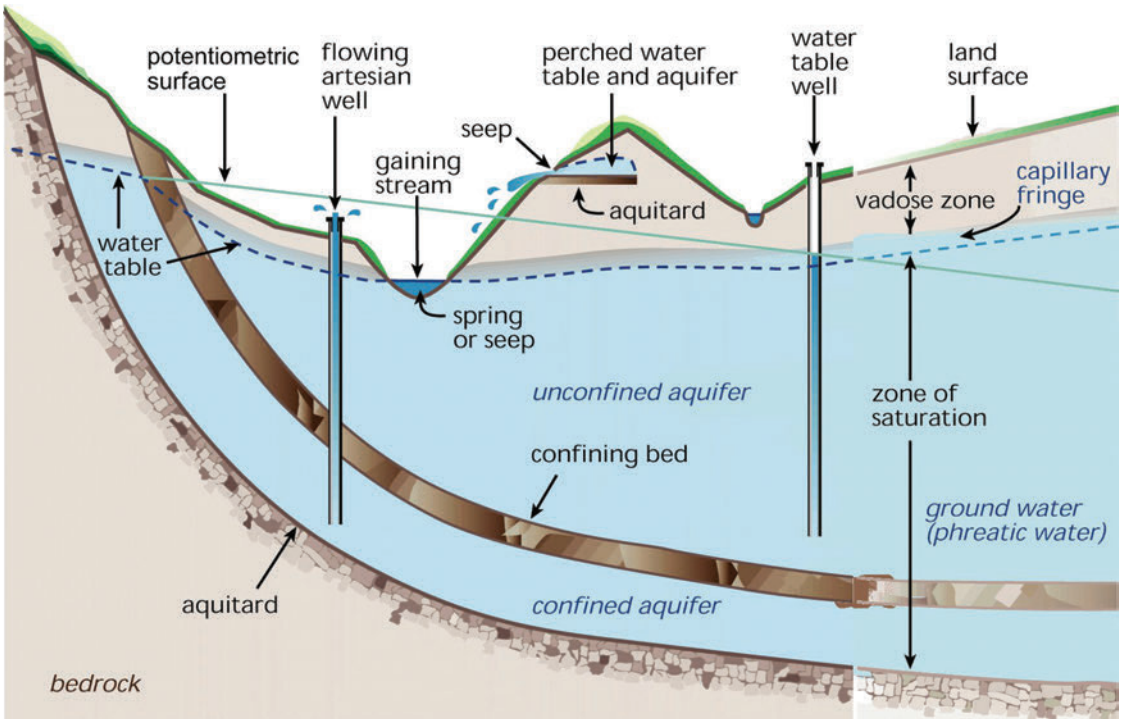


Figure 5.3 Types of aquifers. In an unconfined aquifer the water level in the well is at the water table. In a confined aquifer, the height of water in the well will depend on the amount of pressure within the confined aquifer.

Source: Federal Interagency Stream Restoration Working Group (FISRWG) (1998)

Recharge, storage and discharge

As mentioned in Chapter 1, recharge is the process whereby rainfall eventually finds its way through the vadose zone to the saturated phreatic zone to become groundwater. Once water has reached the water table it is generally not available for evaporation (except through transpiration by the deepest-rooted plants). The area over which this occurs is the *recharge area* for a particular groundwater resource (Figure 5.4), and identifying this, together with establishing the proportion of rainfall that becomes recharge, are important questions for groundwater resource managers to consider.

Groundwater is not only important as a water resource contained in aquifers; flows in most streams are, to a greater or lesser degree, sustained by groundwater flows. Features in the landscapes such as wetlands, ponds and lakes all represent areas where the local or regional water table has intersected the surface. Similarly, a river can be seen as a place where the outward flow of subsurface water gives rise to surface flow. This is referred to as a *discharge zone* (Figure 5.4). It stands to reason that subsurface (i.e. soil and groundwater flows) must be contributing to streamflow or streams would dry up shortly after rain has fallen and flowed away as surface runoff. In Chapter 7, we'll explore this further, particularly from the perspective of what controls the relative proportions of soil and groundwater flows in determining overall flow in a river.

It is traditional to think of groundwater sustaining streamflows during the summer months, which indeed it often does. However, the interaction

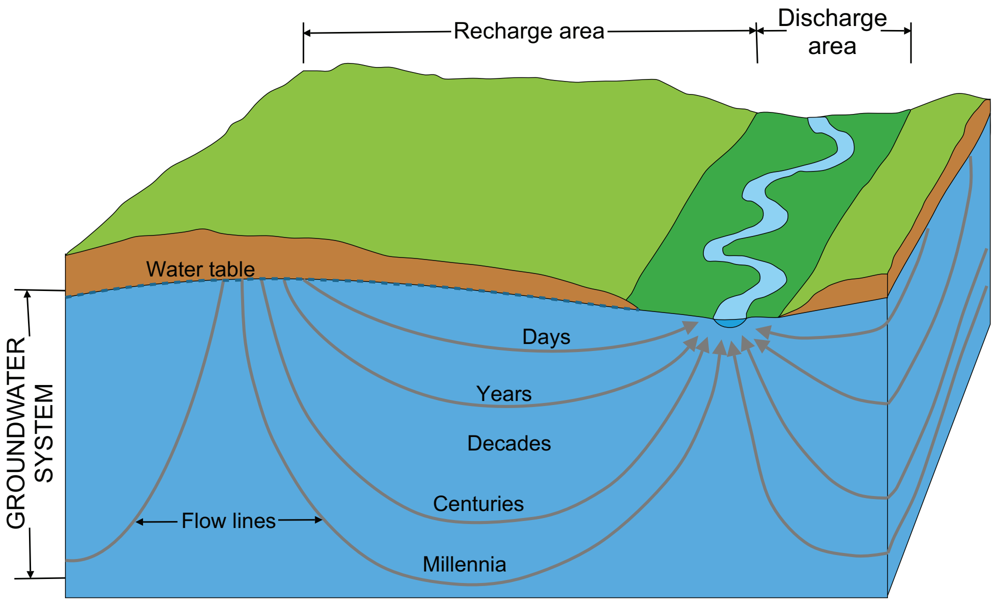


Figure 5.4 Groundwater flow paths from a recharge zone to a discharge zone. Flow paths are of differing length and flow rates, which means that groundwater has a variable residence time in the ground.

Source: Adapted from Skinner et al. (2004)

between groundwater and streamflow is complex and depends very much on local circumstances. Water naturally moves towards areas where faster flow is available and consequently can be drawn upwards towards a stream. This is the case in dry environments but is dependent on there being an unconfined aquifer near to the surface. If this is not the case then the stream may be contributing water to the ground through infiltration. Figure 5.5 shows two different circumstances of interaction between the groundwater and stream. In Figure 5.5(a) the groundwater is contributing water to the streamflow as the water table is high. These are known as **effluent** streams (or **gaining** streams). In Figure 5.5(b) the water table is low and the stream is contributing water to the groundwater. This is commonly the case where the main river source may be mountains a considerable distance away and the river flows over an alluvial plain with the regional groundwater table considerably deeper than stream level. These streams are known

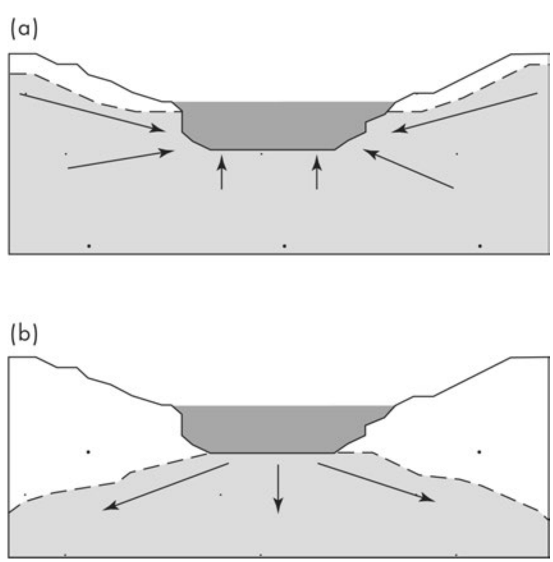


Figure 5.5 The interactions between a river and the groundwater. In (a) the groundwater is contributing to the stream, while in (b) the opposite is occurring.

as **influent** streams (or **losing** streams) and because of this process, may flow only intermittently. The interaction between groundwater and streamflow is discussed further in Chapter 7, especially with respect to stormflows.

In seasonal climates we might expect higher rainfalls in the wet season, and following recharge, more stored groundwater, reflected as a rise in the regional water table. In addition to being a zone of transition between saturated and unsaturated conditions, the water table is also typically seasonally dynamic and may vary more widely and over much longer time frames in relation to rainfall variation and its climatic drivers. We might assume that more groundwater (i.e. a higher water table) might lead to more streamflow, but what is the relationship between these zones of recharge and zones of discharge? How quickly or slowly does groundwater move? We know for instance that similar droplets of water falling near to each other as rain might take very different pathways to reach a stream, one drop taking months or years and the other years or decades (see Figure 5.4). Why does deeper groundwater seem to travel so slowly? What controls the velocity and direction of movement of groundwater? We will consider these issues in the next section.

FUNDAMENTAL PRINCIPLES OF GROUNDWATER MOVEMENT

To make it easier to grasp the fundamentals, we will break this section into two components. Firstly, we will consider what determines the *direction* of groundwater movement. Secondly, we will identify the factors which determine the *rate* at which this occurs (i.e. the speed of movement, or velocity vector if you already know the direction of movement). If one understands something about the rate of movement of groundwater, by implication you also know something about the *residence time* of groundwater. Knowing flow rates, recharge or replenishment rates and residence times are all important considerations in determining the sustainability of any aquifer for water supply. In the Middle East, for example, Saudi Arabia is able to draw on extensive 'fossil water' reserves – so called because they formed in an historic climate characterised by much higher rainfall. These aquifers have had long residence times, but would be characterised by low replenishment rates, and consequently are in a practical sense a finite rather than renewable resource. The case study below provides some insight into how we might determine the age of groundwater.

Case study

METHOD – AGEING GROUNDWATER

An important piece of information for somebody managing groundwater resources is the age of the water contained in the aquifer. This information will give an idea of how quickly any contaminated water may move towards an extraction zone, or how long ago the contamination occurred. Darcy's Law gives an indication of the possible flow rates within an aquifer, but the measurement is at too small a scale to be scaled up to estimate how long it has taken for water to reach a certain position. Frequently there is little idea of where the water has actually come from, so even if you could estimate water velocities you don't know the distance travelled and therefore can't estimate age. In order to overcome this problem, groundwater scientists use the chemistry of different substances dissolved within the water to estimate its age.

Carbon dating is a common technique for the dating of terrestrial deposits but is problematic for young groundwater, since it is only accurate when the sample is more than 1,000 years old. Groundwater is frequently more than 1,000 years old so it is possible to use the technique of carbon dating, looking at the rate of decay of 14C in dissolved organic carbon. Another form of carbon dating looks for 14C resulting from the testing of thermo-nuclear weaponry.

Another dating method, particularly for younger groundwater, is to look for concentrations of materials that we know have been added to the atmosphere by humans. Fortunately for groundwater dating, humans have been very good at polluting the atmosphere with substances that are then dissolved in precipitation. Figure 5.6 shows the concentration of four gases that have been added to the atmosphere in differing amounts. The relative concentrations of these gases in water samples give an estimate for the average age of the groundwater. Tritium, a radioactive isotope of hydrogen with 3 neutrons, was added to the atmosphere in large quantities through explosion of hydrogen bombs, particularly in the 1960s and 1970s. Tritium concentrations reached a peak in 1963 and have since declined to almost background concentrations. Tritium has a radioactive half-life of 12.3 years. Chlorofluorocarbon (CFC) compounds were used in aerosols and refrigeration

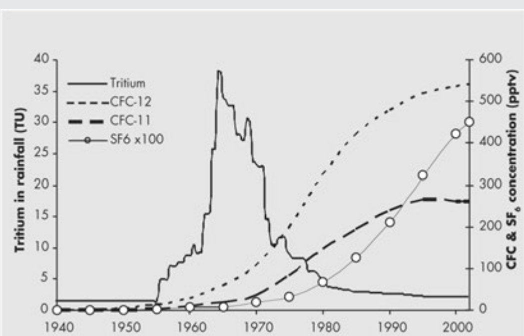


Figure 5.6 Tritium concentrations in rainfall, CFC and SF6 concentrations in the atmosphere 1940–2002. Tritium units (TU) are 1 tritium atom in 1,018 hydrogen atoms. CFC and SF6 are in parts per trillion by volume (pptv).

Source: Figure redrawn from Stewart et al. (2007)

from the 1940s until their banning in the 1990s. CFC-11 concentration has slowly declined since about 1993, while CFC-12 concentration is still increasing, but at a much slower rate than before 1990. Sulphur hexafluoride (SF6) is used for cooling and insulation, particularly in electronics.

Another dating method is to look at the ratio of two isotopes of oxygen and/or two isotopes of hydrogen found in water molecules. When water in the atmosphere condenses to form rain there is a preferential concentration of heavy isotopes of hydrogen and oxygen in the water molecules. The heavy isotope of hydrogen is deuterium (1 neutron) and the heavy isotope of oxygen is 18O. The colder the temperature is at time of condensation the more enriched with deuterium and 18O the water sample will be. In climates with distinct seasons the amount of deuterium and 18O in rainfall samples will vary according to seasons. By taking a series of water samples from rainfall and groundwater, a comparison can be made between the time series. If the groundwater shows considerable variation in deuterium and/or 18O concentrations, then it is relatively recent rainfall. If there is very little variation then it is assumed that the groundwater is a mixture of rainfall from both summers and winters in the past and is therefore older. This is demonstrated in Figure 5.7. The technique of looking at oxygen and hydrogen isotopes is particularly common as a way of determining whether water in a stream is new (i.e. recently derived from rainfall) or old (has been resident in groundwater for some time).

By measuring the concentrations of contaminants like tritium or the ratio of isotopes of oxygen, an estimate of the age of groundwater can be made. The different ways in which water moves through the unsaturated and saturated zones of catchments means that groundwater and streamflows contain water with different residence times. The water in a sample does not have a discrete age, but has a distribution of ages. This distribution is described by a conceptual flow or mixing

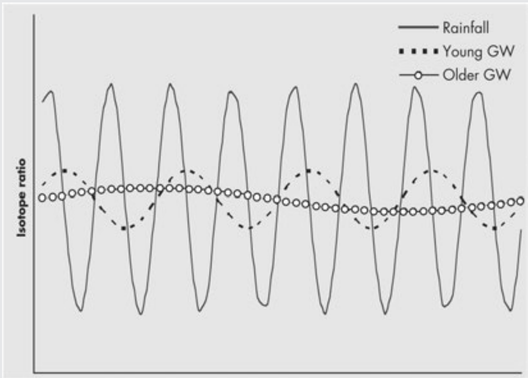


Figure 5.7 Changing ratios of isotopes of oxygen and hydrogen with time in a seasonal climate. Rainfall is heavily influenced by temperature and shows large variation between seasons. The older the groundwater the more dampened down the time series.

model, which reflects the average conditions in the catchment (Stewart et al. 2007). Maloszewski and Zuber (1982) provide an extensive review of these mixing models; in short they account for the size of groundwater reservoir, the concentration of contaminant and the likely time of residence within a well-mixed groundwater reservoir.

Using the techniques outlined here, the average age of groundwater or streamwater can be derived. Studies of groundwater age frequently use a combination of the different techniques to derive an average residence time of water in a catchment, or the groundwater age. It is important to realise that it is an average residence time, not absolute. The water contained in the groundwater reservoir will be a mixture of water that has infiltrated rapidly and some that moved very slowly through the unsaturated zone.

What determines the direction of groundwater movement?

In the introduction to this chapter, we established that in most cases of simple underlying geology, the water table follows the topography, although in a more muted way. This means that at some locations (e.g. below a hill), the water table would be 'higher' than the water table in the valley below. Because of the effects of gravity, we would expect the groundwater at higher elevation to move downwards towards the valley bottom. This of course is true; it is after all the reason why rivers flow downstream from source to sea, and groundwater will also flow 'downwards'. We can express the potential for the river to flow downstream by looking at the differences in elevation. Water held at elevation has **mechanical energy** and therefore the power to do work. For example, the Itaipu Dam on the Brazil/Paraguay border is the world's biggest hydroelectric power scheme. The concrete dam has a height of 196 m (equivalent to a 65-storey building), holding back the waters of the mighty Parana River. From near the top of the wall, water is led into 20 turbines and the combined power of this water is capable of generating 14 GW; providing three quarters of the electricity consumption of Paraguay and 17 per cent of Brazil's. In this case mechanical energy is converted into electrical energy. The reason for this remarkable power output is related both to the naturally high flow rate in the Parana River, but importantly also due to the very large difference in height between the turbine inlets and outlets (118 m). This difference in height means gravity is working in your favour, and is known as **elevation head**.

In the case of groundwater, it is very important to recognise that while elevation head is important, it is not the only factor that is relevant, as Figure 5.8 appears to be suggesting that groundwater can also move upwards against gravity (look at the lower flow paths). To help understand this, take a look at Figure 5.8. In the top figure, if the tap is opened, water will clearly flow down from the tank and out through the tap. This is because of the action of gravity and the elevation head that exists. Now consider the lower figure. Here a bend is present in the pipe, which means that in order to escape, water needs to flow upwards against gravity.

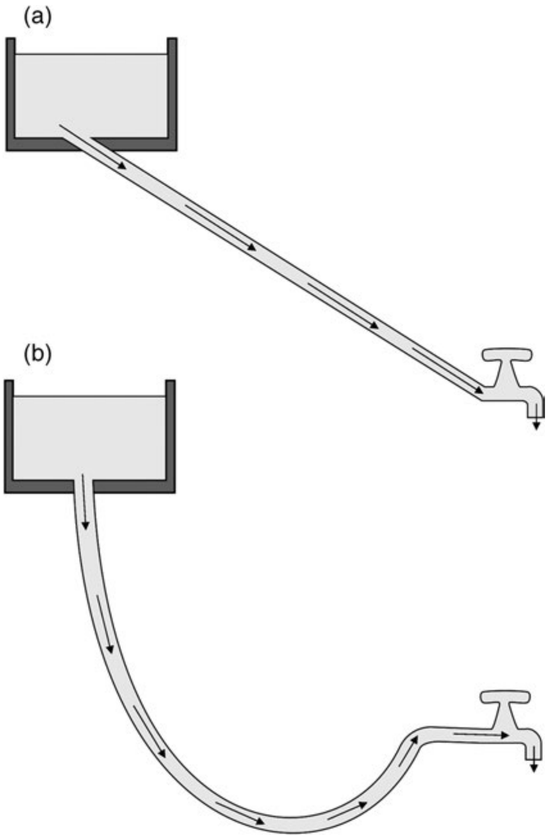


Figure 5.8 Can water flow upwards? An analogy using a simple header tank and hosepipe. (a) Straight hosepipe in which flow is always downwards. (b) Hosepipe with a loop in which the water must always flow upwards to reach the tap. The laminar nature of most groundwater flow means that it tends to behave more as if it was in a stack of pipes like this, rather than like water flowing down a stream channel.

Source: Reproduced from Younger (2007), with permission

We know from experience that if we did open the tap in the lower figure, water would certainly still flow out, despite having to flow 'uphill'. This suggests that it is not elevation head alone that influences groundwater movement.

We have already established that groundwater can be at pressure (see earlier reference to artesian wells, and also the concept of pore pressure). The fact that it is at pressure means it has the ability to move, independent of elevation, in other words it has a **pressure head** too. In some cases, the pressure head means that groundwater can flow uphill (rather than just downwards because of the elevation difference). It follows that if we are to understand the ability of groundwater to move, we need to consider both the *pressure head* and the *elevation head*. Importantly, it is the combination of these two heads, or the **hydraulic** or **total head**, that is critical in determining the direction of groundwater movement.

Figure 5.9 illustrates this a little more clearly, showing that groundwater flows from high total head to low total head, rather than just from points of high pressure to low pressure, or high elevation to low elevation. Spend a few minutes looking at these figures and the accompanying explanation.

The concept of a fluid having mechanical energy was formalised in 1738 by Daniel Bernoulli (1700–1782). Bernoulli's law is a fundamental equation of fluid mechanics, and recognises firstly that that objects have three types of mechanical energy; kinetic, potential and pressure, and secondly that the sum of these remains constant (the idea of

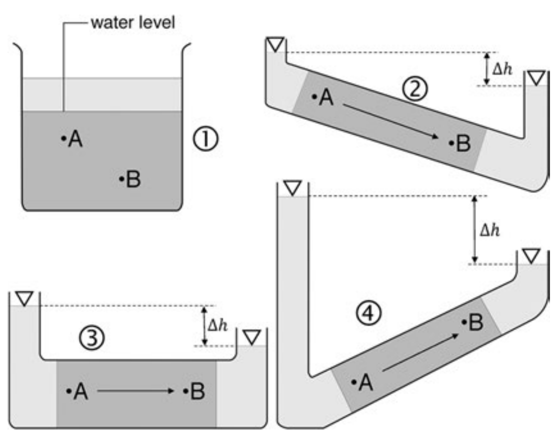


Figure 5.9 Groundwater flows from high total head to low total head, rather than from just high to low pressure or high to low elevation.

Source: Reproduced from Younger (2007), with permisssion *conservation of energy*). Pioneering work by Hubbert (1940) in applying these concepts to groundwater laid important foundations for the study of groundwater movement. He recognised that groundwater flow occurred due to spatial variation in the 'mechanical energy content per mass of water', and coined the term 'potential' to represent this. This is the origin of the term potentiometric surface shown in Figure 5.2 – the potentiometric surface is an imaginary line showing the height to which water would rise if wells intercepted a confined aquifer. His modification of Bernoulli's equation shows that the *hydraulic head* (or *total mechanical energy* available) is the sum of three components – the first term is the *elevation head*, the second term is the *pressure head* and the final term the *velocity head* (Equation 5.3).

$$h = z + \frac{P}{\rho_w g} + \frac{v^2}{2g} \tag{5.3}$$

Where h is the hydraulic or total head (mechanical energy available), z is the elevation head, P is the pressure exerted by the water column, ρ_w is the fluid density, v is the velocity and g is gravitational acceleration.

The last term in the equation above represents the kinetic energy component, or the energy associated with movement. Because it is very small in groundwater it is ignored, but as we will see in a later chapter, is very important in river flow. So in groundwater, total head can be expressed as Equation 5.4:

$$h = z + \frac{P}{\rho_w g} \tag{5.4}$$

Those of you that have recently done a school science module might recognise something familiar about the second term, the pressure head. How do you calculate the pressure under a column of water? This is known as the hydrostatic pressure and is given by Equation 5.5:

$$p = h\rho_w g \tag{5.5}$$

Where p is the pressure in the fluid, h is the height of the fluid column, and as before, ρ_w and g are the density of the fluid and gravitational acceleration, respectively. If you re-arrange the terms of the equation, solving for h, rather than p, what do you get? Clearly, this is equivalent to the second term in Equation 5.4. In other words, the pressure head term in Bernoulli's equation is really just the height of the water column. So it is the height of this water column and its elevation above a datum that determines the total head (Figure 5.10).

Although it is important to realise that total head is the combination of pressure head and elevation head, in practice we do not have to calculate them separately. Total head is relatively easy to measure in the field (Younger 2007). We'll discuss measurement techniques a little later in this chapter, so for now let's just assume that we have either measured the depth to the water table in Figure 5.10, or alternatively have used a pressure transducer to measure the pressure, P, at the point noted in the well and calculated the height of the column using the re-arranged hydrostatic equation.

The value of doing this is that once the total head can be calculated at two or more points, the

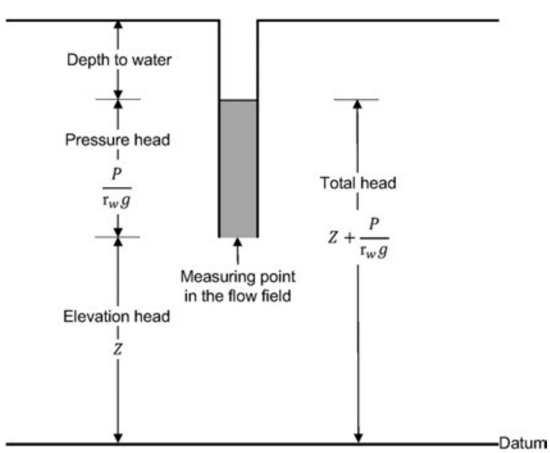


Figure 5.10 Relationships between total or hydraulic head, pressure head and elevation head.

direction of groundwater flow can be determined. This is straightforward, as groundwater always moves from a high total head to a low total head.

What determines the rate of groundwater movement?

We will now turn to considering the factors that determine the speed of groundwater movement. The pioneering work in this respect was undertaken by a nineteenth-century French engineer, Henry Darcy. Darcy was concerned with the water supply for Dijon, and in particular how impurities from aquifer-fed spring water could be removed by sand filtration (Darcy 1856). Darcy designed a series of observations on the characteristics of flow through sand, observing that the 'rate of flow of water through a porous medium was proportional to the hydraulic gradient' (Darcy 1856). The **hydraulic gradient** being the difference in elevation between two points, divided by the distance between them. In other words, and applied in a groundwater context, the slope of the lines of hydraulic head. Linking back to the previous section, while the *direction* of groundwater flow between two or more points is determined by the direction of this slope (high total head to low total head), the *rate* at which this occurs is proportional to the hydraulic gradient between these points (Figure 5.11). All other things being equal, the higher this slope, the greater the flow. The concept of hydraulic gradient thus includes the ideas of elevation and pressure head discussed earlier. In an unconfined aquifer it can be assumed that the hydraulic gradient is equal to the drop in height of water table over a horizontal distance (i.e. the elevation head). In a confined aquifer it is the drop in phreatic surface (i.e. the level that water in boreholes reaches given the pressure the water is under) over a horizontal distance.

As you will recall from your earlier science education, to change a relationship from one of proportionality to an equation, a coefficient of proportionality is required. Darcy's work found that this coefficient of proportionality was strongly related to the grain size of the sand, in other words, largely a property of the medium through which the water is moving. This factor is now known as the **hydraulic conductivity**, sometimes also referred to as the coefficient of permeability. There are many different ways of formulating Darcy's law, but the two most

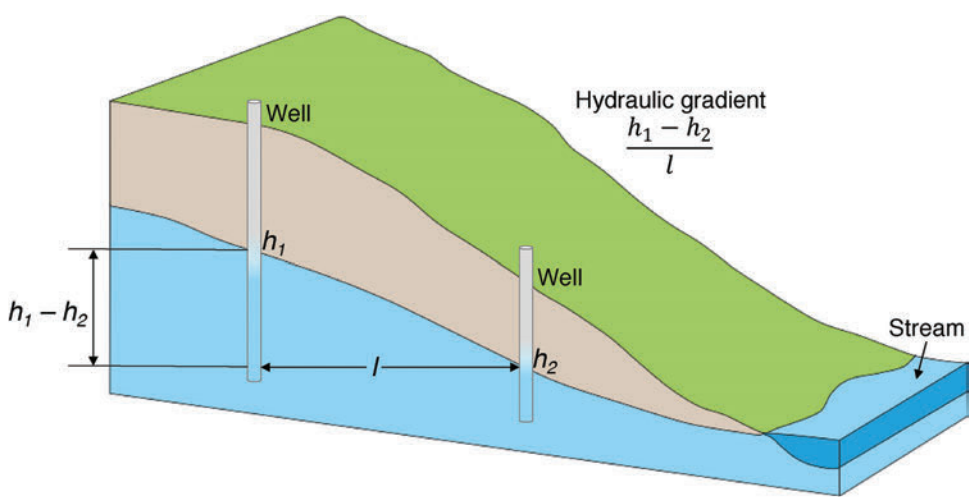


Figure 5.11 The concept of hydraulic gradient.

common and easily understood forms are shown in Equations 5.6 and 5.7.

$$v = K_{sat} \cdot \left[\frac{(h_1 - h_2)}{L} \right] \tag{5.6}$$

Where v is the velocity of groundwater flow (m/s), K_{sat} is the saturated hydraulic conductivity (m/s i.e. the same units as v), h_1 and h_2 are the heights of the water table (m) at two locations and L is the distance between them (m). The term in square brackets is the *hydraulic gradient*. Groundwater hydrologists are not only interested in the rate of flow, but also the groundwater discharge – or the volumes of flow over a given time period. This is obviously important because usually we want to be able to extract volumes of groundwater. In this case we need to think about the 'area' over which this flow is occurring. Think about this as the cross-sectional thickness of the aquifer, at right angles to the direction of flow. In this case, Darcy's equation can be written as:

$$Q = A \cdot K_{sat} \cdot \left[\frac{(h_1 - h_2)}{L} \right] \tag{5.7}$$

Where Q is the *groundwater discharge* (m^3/s), A is the *cross-sectional area* (m^2) through which the flow occurs, and the remainder is the same as Equation 5.6.

Darcy's law is an empirical law (i.e. based on experimental observation) that appears to hold under many different situations and spatial scales. It underlies most of groundwater hydrology and is very important for the management of groundwater resources. Groundwater models such as the US Geological Survey's (USGS's) MODFLOW 6 (Langevin et al. 2017), essentially solve Darcy's equation in three dimensions, so saturated hydraulic conductivity needs to be known in the vertical direction (z), as well as over the horizontal planes (x,y).

PROPERTIES OF AQUIFERS

Hydraulic conductivity

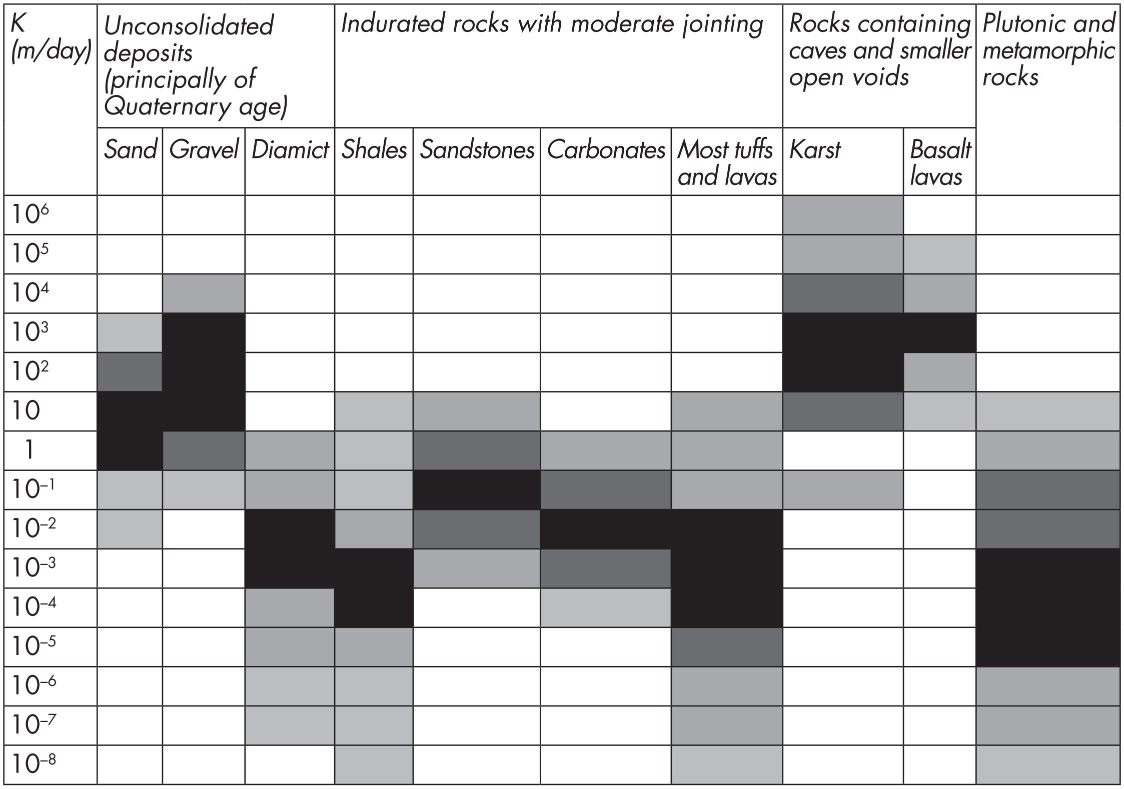
We've seen from the earlier section that establishing the direction of groundwater flow and the hydraulic gradient can be relatively straightforward – all that is required is observations from two or more boreholes. However, calculating corresponding velocities or discharges is much more challenging because this requires you to know the *saturated hydraulic conductivity* (K_{sat} in Equations 5.6 and 5.7), and is a fundamentally important parameter in groundwater studies. Broadly speaking, this is the ability of a porous medium to transmit water, and is a consequence of both the *size* of pores within the soil or rock and the *interconnectivity* between these pores. Clearly, it is related to the idea of *effective porosity* mentioned earlier. While effective porosity is simply the ratio of connected pores to the total volume of material (dimensionless), the saturated hydraulic conductivity is a measure of the rate at which water can flow through this material and has units in distance per unit time (e.g. m/s, or more typically m/day because of the slow rate of movement).

Table 5.2 gives typical values of saturated hydraulic conductivity for a range of material types. Note the wide variation in these values; in fact, hydraulic conductivity is known to vary over 13 orders of magnitude. As we shall shortly see, hydraulic conductivities also vary spatially at both micro and macro scales, representing a further challenge for groundwater scientists.

Although in the above discussion, hydraulic conductivity is considered a property of the material, this is not entirely true. It is also a property of the fluid passing through – particularly its density and viscosity. But if we assume that we are considering just freshwater, then this is not a problem. However, if we are concerned with *saline* groundwater, which has different density and other properties, we would need to make further adjustments.

A further complicating factor is that hydraulic conductivity may well vary depending on which direction it is measured in (e.g. horizontally or vertically). This is not surprising since the structure

Table 5.2 Typical values of saturated hydraulic conductivity (K_{sat}) for different types of material



Source: Reproduced from Younger (2007)

of the rock (e.g. bedding planes, orientation of particles) is likely to be different. A completely uniform material would be *homogeneous* (K_{sat} is the same everywhere in the aquifer) and *isotropic* (K_{sat} is the same in all directions through the aquifer). In reality, many aquifers are the opposite, being both *heterogeneous* (K_{sat} varies spatially through the aquifer) and *anisotropic* (K_{sat} varies in different directions such as parallel vs perpendicular to the bedding).

Although K_{sat} can be measured from a small sample in the laboratory (Klute and Dirksen 1986), in the management of water resources it is more common from larger-scale well-pumping tests (see Freeze and Cherry (1979) for more details). The well-pumping test gives a spatially averaged K_{sat} value at the scale of interest to those concerned with water resources, and once established for a particular region of interest can be applied in Darcy's equation to establish groundwater flow rates.

Specific yield and specific retention

Earlier in this chapter we introduced the concept of *porosity*, or the proportion of a rock that is comprised of pores. We clarified this by also referring to the idea of *effective* porosity, where it is only the connected pores that are relevant. We need to modify our understanding further by considering whether all the water stored in connected pores would be available for abstraction. The answer to this question is that it is not, and the reason for this relates to **capillary forces**. Capillary action occurs when water interacts with a solid material. You will be familiar with the idea of the level of a soft drink inside a straw rising above the level of the drink in the container when the straw is placed in the liquid. The liquid is sucked up the straw and if you looked closely you would also see **adhesion** on the edges – where the liquid level curves upwards around the straw (a concave meniscus). If you placed straws of various diameters in the drink you would also see a difference; the smaller the diameter of the straw the higher the level the liquid in the straw would reach. Capillary action is a suction (i.e. a force) that acts against gravity and is a characteristic of the physics of small spaces. It is primarily a consequence of the **surface tension** property of water. In the natural environment, droplets of water form because of the relatively high surface tension of water, arising from the cohesive properties of water. **Cohesion** means that liquid molecules have greater attraction to one another as opposed to the air molecules (refer back to the dipolar nature of water discussed in Chapter 1). But when in contact with some surfaces, water molecules have a stronger attraction to the molecules of the surface than each other (or the air) – this is *adhesion*. For example, on mineral surfaces water tends to spread, wetting the surface. In our other example, the liquid in the straw is more attracted to the surface of the straw and this is why there is a concave meniscus in the straw. In small spaces such as soil and rock pores, surface tension (caused by cohesion within the liquid) and adhesive forces between the liquid and the solid particles act together to move water against gravity. Importantly these forces also act to hold water against gravity. As was the case with the straws of differing diameter, these forces are more pronounced the smaller the spaces. Since there is a relationship between the size of grains and the pores that are formed between them, this means that there is a general relationship between grain size and capillarity or the strength of this suction; the finer the grain size the stronger this force.

Now let's consider what the practical implications of capillary forces are for groundwater. Firstly, you will remember from Figure 5.1 that we referred to the *capillary fringe* in the vicinity of the water table. Now you will be able to better understand that this occurs because water is drawn upwards from the saturated zone into the unsaturated zone by these capillary forces. The width of this zone will depend on the grain size and will be much lower for say, gravels than fine grained mud deposits.

The second practical implication relates to the question we posed at the start of this section – is all water that might be contained within interconnected pores of rock available to us? We said earlier that it wasn't because of capillary forces. These forces act to hold water back and can be stronger than gravity – this means that water will not flow to the point where we might abstract it from. This gives rise to two inter-related properties of an aquifer, **specific yield** (normally symbolized as S_y) and **specific retention**. *Specific retention* is the water that is held back in the aquifer because of capillary forces. Conversely, *specific yield* is the water that can be extracted from the aquifer. It follows logically that the sum the *specific retention* and the *specific yield* must equal the *effective porosity*. Younger's (2007) representation of this is very helpful in drawing out these distinctions (Figure 5.12).

The relationship between the concepts of porosity, specific yield and specific retention, for unconsolidated deposits can be seen in Figure 5.13. As the particle size increases, both the porosity and the specific retention decrease. Although clay has the highest porosity, the strength of capillary forces is highest and so has a low specific yield. In this example, sand has the highest specific yield and therefore would be the most useful aquifer.

Storativity

The concept of specific yield and specific retention relate only to unconfined aquifers, where the water table can move under the influence of gravity and in response to say, pumping. In a confined aquifer the pores do not 'drain' under the influence of gravity and the factors controlling storage behaviour are very different (see Younger (2007) for a more complete description of these). The term **storativity**, (usually symbolised as S), is used in this case. This is defined as 'the amount of water which can be removed from a unit volume of confined aquifer per unit decline in water level (measured in wells penetrating that aquifer)' (Younger 2007: 17).

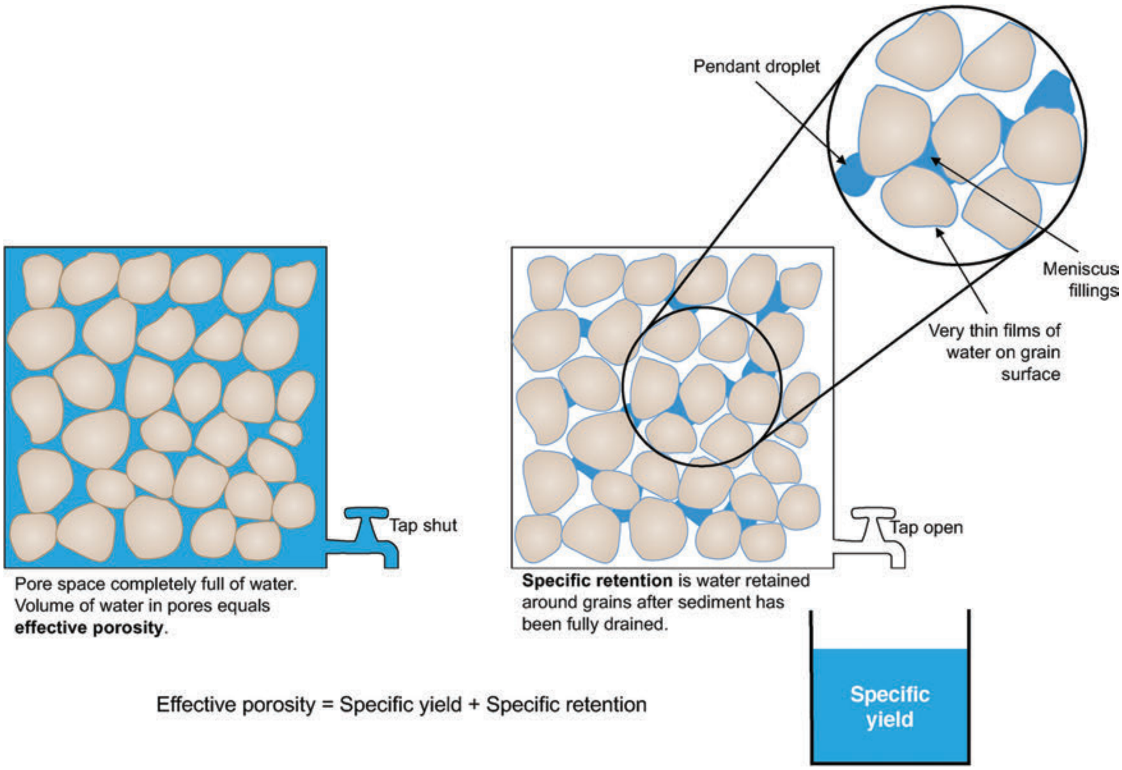


Figure 5.12 A conceptual sketch to explain the storage properties of unconfined aquifers.

Source: Reproduced from Younger (2007), with permission

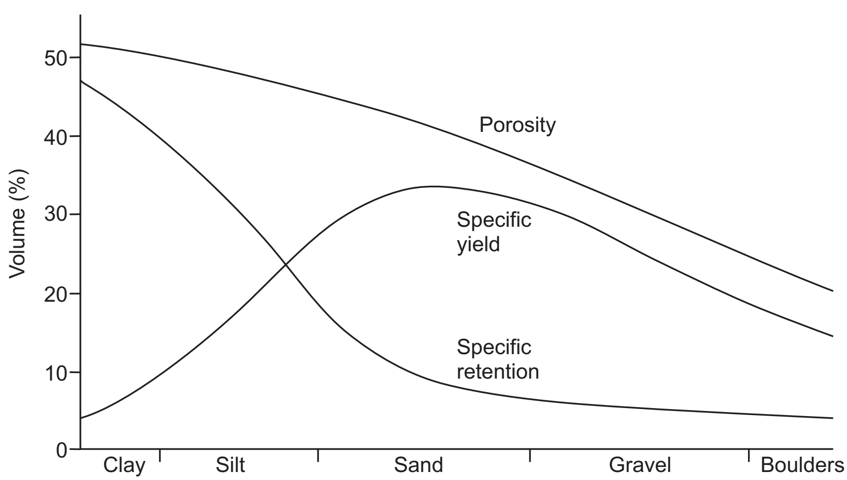


Figure 5.13 The relationship between porosity, specific yield and retention for different types of consolidated material.

Source: Reproduced from Ward and Robinson (2017), with permission

Transmissivity

Transmissivity extends the idea of hydraulic conductivity by considering the thickness over which this process is occurring. In a way, it is an indication of how useful the aquifer might be from a groundwater resource perspective. For example, a thin layer of sediment might have a very high hydraulic conductivity but as it is a thin layer this might not yield much water. In contrast a much thicker layer – which might have a much lower saturated hydraulic conductivity – could yield much more water, just because it is thicker. Transmissivity is estimated by Equation 5.8.

$$T = K_{sat} \cdot b \tag{5.8}$$

Where T is transmissivity (typically m^2/day), and K_{sat} is saturated hydraulic conductivity (m/day) and b is the saturated thickness of the aquifer (m). Note though, that b can vary over time, so this is a dynamic rather than static property of an aquifer. This formulation also assumes that K_{sat} does not vary significantly over the aquifer thickness. The many different terms used in groundwater hydrology at first appear confusing, but it is important to understand them as they all refer to specific properties of the way water moves, or doesn't move, underground.

MEASURING AND ESTIMATING GROUNDWATER

Wells and piezometers

The main measurement techniques for water in the saturated zone are through **wells** and **piezometers**. The term **borehole** is often used to denote a well that has been drilled for the purpose of extracting water. Both wells and piezometers measure the height of the water table but in slightly different ways. A *well* is drilled into the ground using specialised equipment, and usually a tube surrounded by some form of casing is inserted. The latter could be slotted steel or plastic and is sometimes called a *well screen*. This casing prevents the sides of the well closing in, but importantly is permeable. This allows water to percolate through from all parts of the column. The well screen can also be surrounded by a gravel pack which acts as a filter, particularly if the well is for water supply. Wells are commonly used for water extraction and monitoring the water table in unconfined aquifers. Where the purpose is monitoring, water depth can be estimated by use of a *dipper* (Younger 2007). This is a measuring tape which contains a pair of electrodes. The tape is lowered into the borehole and when the electrodes comes into contact with the water, an electrical circuit is completed and a sound alarm and/or bulb is illuminated. This allows the depth to the water table to be read off the tape. The hydraulic head can then be established by subtracting this depth from the surveyed ground elevation at the top of the well. Where frequent or continuous measurements are required a pressure transducer, attached to a data logger can be installed. This can be set to take measurements at specific times, or alternatively when user-selected thresholds of change in head are detected (Younger 2007). Alternatively, head can be estimated and logged by use of a pressure transducer. This is a submergible sensor normally placed below the lowest anticipated groundwater level, and works on the basis of measurements of deflection of a membrane exposed to pore water pressure on the one side and atmospheric pressure on the other (Fitts 2013). The depth that the pressure transducer is installed at needs to be known (refer back to Figure 5.9) in order for the hydraulic head to be calculated using Equation 5.4.

A *piezometer* is a special type of well that is designed to measure the hydraulic head at a particular location and depth. The principle of construction is the same, except that the tube is narrower (usually <30 cm), and the area of perforation of the tube is only in the vicinity of its base. Usually the borehole is also filled with a material such as bentonite to prevent water from the surface of or anywhere else along the borehole seeping in. This means that the height to which the water rises in the tube (or piezometer) is thus a record of the mechanical energy exerted by the groundwater at the base of the tubing. Alternatively, as in the case of wells, a pressure transducer can be used. Piezometers are therefore used to measure the total head at different depths in the aquifer, sometimes as ‘nests’ within the same borehole to see variation with depth or alternatively across an aquifer to establish spatial variation. With sufficient point observations, and via interpolation between these points of known pressure head, lines of **equipotential** can be drawn. Think of these lines (usually represented by the symbol Φ), as lines of equal head, much like contours represent topography, or isohyets represent lines of equal rainfall (see Chapter 2). The interpolated 2-D surface is the **potentiometric surface** (see Figure 5.2), also called the **piezometric surface**. Once this surface is known, spatial patterns of groundwater flow can be established, as groundwater flow will occur in **streamlines** in a direction perpendicular to the lines of *equipotential*, just in the same way as flow on the land surface occurs perpendicular to contour lines.

Once one, or multiple piezometers or wells are installed, a variety of piezometer and pumping tests can be used to help determine the characteristics of the aquifer under investigation. Review of these is beyond the scope of this chapter, but most standard groundwater texts will explain these.

Surprisingly, groundwater (along with other terrestrial storage) can also be estimated from space by satellites. Becker (2006) gives a useful overview of the potential for satellite remote sensing of groundwater. Some approaches are only applicable at the scale of the very large basin, but nevertheless form a useful addition to the range of techniques for tracking total terrestrial storage over time. The case study below provides a little more information on this approach.

Case study

MEASURING WATER STORAGE FROM SPACE

Through remote sensing technologies it is possible to monitor change in Earth's water storage. The Gravity Recovery & Climate Experiment (GRACE) was a revolutionary satellite mission launched in 2002. Operated jointly by NASA and the German Aerospace Center, the mission originally had a lifespan of 5 years. However due to its major success, the mission was extended until 2017 (with the satellites falling out of orbit in early 2018).

The GRACE mission consisted of two satellites in tandem orbit (≈500 km above the Earth's surface). The on-board instruments constantly measured the distance between the two satellites which fluctuated around 200 km as a result of Earth's changing gravitational field. This idea of the two satellites 'chasing' one another gave the pair their nicknames of 'Tom and Jerry' amongst ground teams. The distance measurements were reported with 1 μm accuracy. As the instruments were passive sensors, data produced was not limited in terms of cloud cover or solar illumination, unlike many optical satellite instruments.

Monthly measurements from the satellites were used to produce mathematical representations of the Earth's gravity field, with a spatial resolution of 380 km. The main drivers of such changes are shifting oceanic and atmospheric circulations and the redistribution of water within the hydrological cycle (e.g. changing ice sheet mass). However, the monthly temporal resolution of GRACE datasets has been noted as a limitation of the mission, especially when compared to traditional hydrological datasets.

Forming the monthly datasets are ≈7,500 individual gravity profiles (≈250/day). These profiles can be directly related to water storage. GRACE ground teams used the gravitational profile to quantify monthly anomalies in the Terrestrial Water Storage (TWS) to an accuracy of 1.5 cm. TWS represents water stored at all levels – groundwater, soil moisture, surface water, snow, ice, and vegetation biomass. The GRACE mission provided the only satellites able to monitor water storage below the first few centimetres of the surface. However, GRACE TWS data has no vertical resolution and therefore users cannot easily distinguish between water stored in the different vertical elements of the system.

In the field of hydrological science GRACE has led to a number of key advances. The key to GRACE's success in this area is that change in TWS (ΔTWS) is a commonly used parameter in water budget equations (Rodell et al. 2004; Swenson and Wahr 2006). GRACE has been used in a range of applications such as monitoring the discharge of large river systems (Syed et al. 2005) and estimating changes in the mass of the Greenland and Antarctic Ice Sheets (Luthcke et al. 2008). Further, while the quantification of different water storage components from GRACE TWS data is challenging, some attempts have been successful. For example, monitoring of surface water variations (Han et al. 2009), and the estimation of groundwater depletion due to irrigation (Voss et al. 2013). A number of studies have used GRACE data to evaluate long-term groundwater changes in various regions, for example, India (Chen et al. 2014; Chinnasamy et al. 2015), Canada (Huang et al. 2016; Hachborn et al. 2017), China (Chao et al. 2016) and in the USA (Texas) (Long et al. 2013). Lakshmi (2016) provides a good overview of possible future directions in using satellite data for groundwater investigations.

A follow-on mission named GRACE-FO (Follow-On) launched on 22 May 2018. GRACE-FO will continue to monitor changes in Earth's gravitational field in the same way as its predecessor.

GRACE data (2002–2017) can be accessed and viewed through the JPL GRACE data portal (http://grace.jpl.nasa.gov/data/get-data/).

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ESSAY QUESTIONS

1. **Compare and contrast the terms *aquifer, aquitard, aquiclude,* and *aquifuge* and explain why these are relative rather than absolute terms.**
2. **Define the terms *hydraulic head* and *saturated hydraulic conductivity* and explain their importance in understanding groundwater flow.**
3. **Explain the terms *confined* and *unconfined* with respect to aquifers and describe how artesian wells come about.**
4. **Explain the fundamental principles controlling the direction and rate of groundwater flow.**

FURTHER READING

Anderson, M. (2008) *Groundwater*. IAHS Benchmark Papers in Hydrology Series. IAHS Press, Wallingford.
Forming part of the IAHS 'benchmark paper' series, this is a compendium of historic papers and associated commentary reflecting the evolution of understanding in the field.

Freeze, R.A. and Cherry J.A. (1979) *Groundwater*. Prentice-Hall, Englewood Cliffs, N.J.
A classic text on groundwater (including soil water).

Hiscock, K. and Bense, V. (2014) *Hydrogeology: Principles and practice*. (2nd edition). Wiley Blackwell, Chichester.
A thorough, advanced text.

Kendall, C. and McDonnell J.J. (eds) (1998) *Isotope tracers in catchment hydrology*. Elsevier Science, Amsterdam.
Detailed book on groundwater and streamwater ageing techniques.

Price, M. (1996) *Introducing groundwater* (2nd edition). Chapman and Hall, London.
An introductory text on groundwater.

Robinson, M. and Ward, R. (2017) Groundwater. In: *Hydrology: Principles and processes*. IWA publishing, London.
An excellent read for a little more depth without having to move to a whole text.

Younger, P. (2007) *Groundwater in the environment: An introduction*. Blackwell Publishing, Malden, M.A.
An excellent introductory text written in a very readable style.

6

SOIL WATER

LEARNING OBJECTIVES

When you have finished reading this chapter you should have:

- An understanding of the fundamental forces influencing the behaviour of water in the soil.
- A knowledge of how the properties of soil influence water retention characteristics.
- An understanding of the principles of water movement in the soil and how the soil absorbs, retains and releases water, thereby regulating runoff.
- A knowledge of the techniques used to measure infiltration.
- A knowledge of the techniques used to measure and estimate the amount of water stored as soil moisture.

WATER IN THE UNSATURATED ZONE

In the previous chapter we considered water below the water table in the saturated zone. For reasons which will become evident later in the chapter, the behaviour of water in the **unsaturated zone** is a little more complicated. The unsaturated zone is also variously referred to as the **zone of aeration**, or the **vadose zone**. We should also point out at this stage that by referring to it as the unsaturated zone, this does not mean that it is never saturated, only that its typical condition is one of being less than saturated. Water in the unsaturated zone is called **soil water** or **pore water**.

Since the majority of water in the unsaturated zone is held in the soil, we begin this chapter by developing an understanding of the composition and structure of soil, including the fundamental forces that determine the behaviour of water in the vadose zone. We then move on to developing an understanding of how some of the hydrological properties of a soil can be represented by parameters, and how from these we can characterise how soil redistributes water.

CHARACTERISTICS OF SOILS

Soil is essentially a continuum of solid particles, air and water. A completely dry soil consists of solid particles with air in the spaces between the particles (the voids), and no water. In contrast, a soil is saturated when all the voids are filled with water and there is no air, as it has been displaced. The movement of water in a soil therefore depends on the processes of filling and emptying the pore spaces and the various forces that come into effect at this micro-scale. Before we discuss these, we will consider the particle component of soils and how these determine the overall hydrological characteristics of a soil. We also need to consider how soils are formed and evolve as this determines how they change across the landscape, which in turn determines in part how hydrological processes vary across the landscape.

Soil composition and texture

Some soils (**mineral soils**) are comprised almost completely of just mineral particles – the tiny fragments of rock they were formed from. Similarly, these parent rocks are formed from a range of minerals (e.g. quartz, feldspar) each in turn having their own characteristics such as shape, hardness and weathering properties. Clearly the underlying geology will therefore be a key determinant of the type of soil present in an area. However, we also need to remember that soils are formed through deposition, so may also reflect the upstream geology as rocks and minerals weather, erode and are transported downstream. Other soils, especially those which develop in heavily vegetated areas, have a high component of organic matter (e.g. 12–18 per cent), and these are known as **organic soils**. In reality this is a continuum and most soils will have some organic matter (typically between 1 and 6 per cent) in the upper layer. This occurs because of the breakdown of vegetation and will be higher where there is a deeper litter layer (e.g. forests). The presence of organic matter in a soil is important for a range of biological processes (e.g. it is an important source of energy for bacteria involved in denitrification), and it also influences some hydrological processes we will discuss later (e.g. infiltration). From here on, we will focus on mineral soils.

One of the most critical factors determining the behaviour of water in a soil is the grain size of the soil particles. Since soil is seldom made up of a uniform size of particle, it is more correctly the *grain size distribution* of the particles. In other words, what proportion of the soil is made up of particles of different sizes. Unfortunately, there is not a standard system for classifying the size of particles, agronomists, for example, have one approach while soil scientists and geotechnical engineers have another (Figure 6.1). The reason for this is that they each focus on a slightly different property of a soil, and for different purposes. Where they differ is in relation to the slightly larger particles; most systems agree that any particle below 0.002 mm (2 μm) is a clay. Clays – and there are many different types – are minerals in their own right and are formed as products of the weathering process. For example, the mineral feldspar, which is a common mineral found in rocks such as granite, weathers into clays. Particles above 2 to 6 μm are considered sands, and these might be comprised of small fragments of minerals such as feldspar and quartz. The coarse sand fraction is large enough to be seen and felt when rubbed between the fingers and thumb, it has **texture**. Intermediate between a clay and a sand particle is silt. Individual particles of silt and clay can't be felt but clays tend to make the soil feel sticky, whereas silt particles impart a silky and sometimes slightly sticky feel (White 2006). We use the term soil **texture**, to refer to the grain size properties of a soil, and in a way it is short hand for 'grain size distribution'.

Earlier we said it is the *grain size distribution* which has an important bearing on the hydrological behaviour of the soil, in other words, the relative proportion of sand, silt and clay in a soil exerts a very strong influence on the ability of a soil to absorb, retain and release moisture. In fact, the relative proportions of these not only determine the hydrological behaviour of a mineral soil but also its textural classification, often used as the 'name' or first description of the soil by those who are not soil scientists. Figure 6.2 shows examples of triangular diagrams which classify a soil into one of several

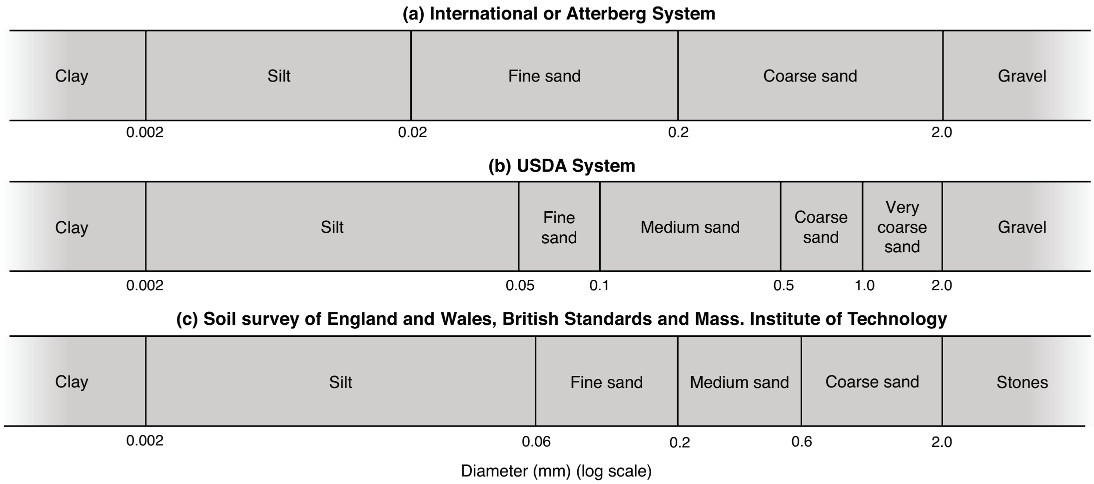


Figure 6.1 Different approaches to classifying soil particles: (a) the International or Atterberg system; (b) the United States Department of Agriculture (USDA) system; and (c) the system used in the soil survey of England and Wales, British Standards and by the Massachusetts Institute of Technology.

Source: Redrawn from White (2006)

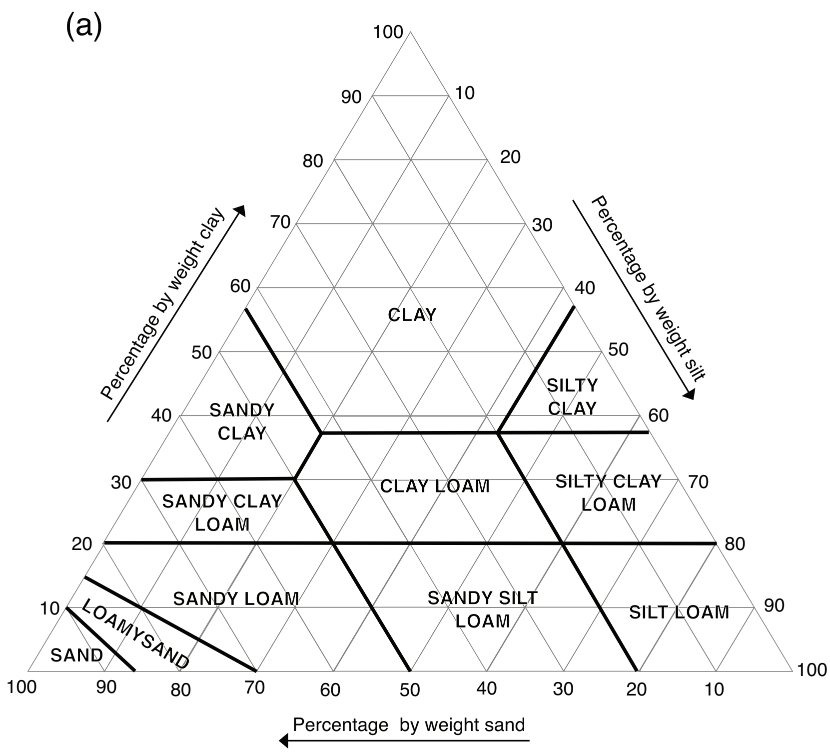
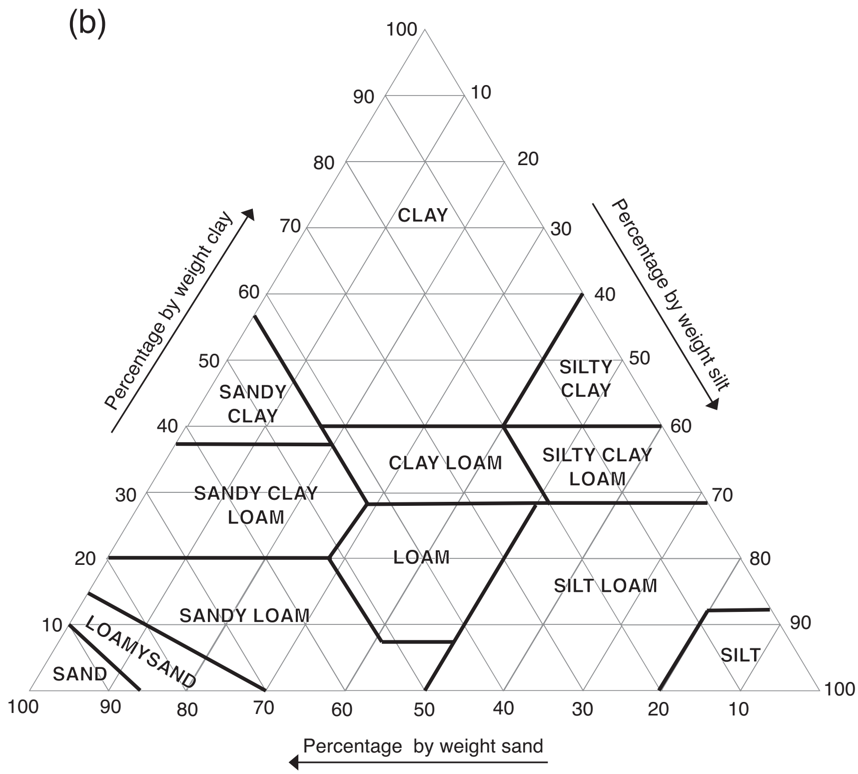


Figure 6.2 Examples of textural triangles from (a) United Kingdom, (b) United States and (c) Australia for classifying soils into their textural classes.

Source: Adapted and redrawn from White (2006) and Dingman (2002)



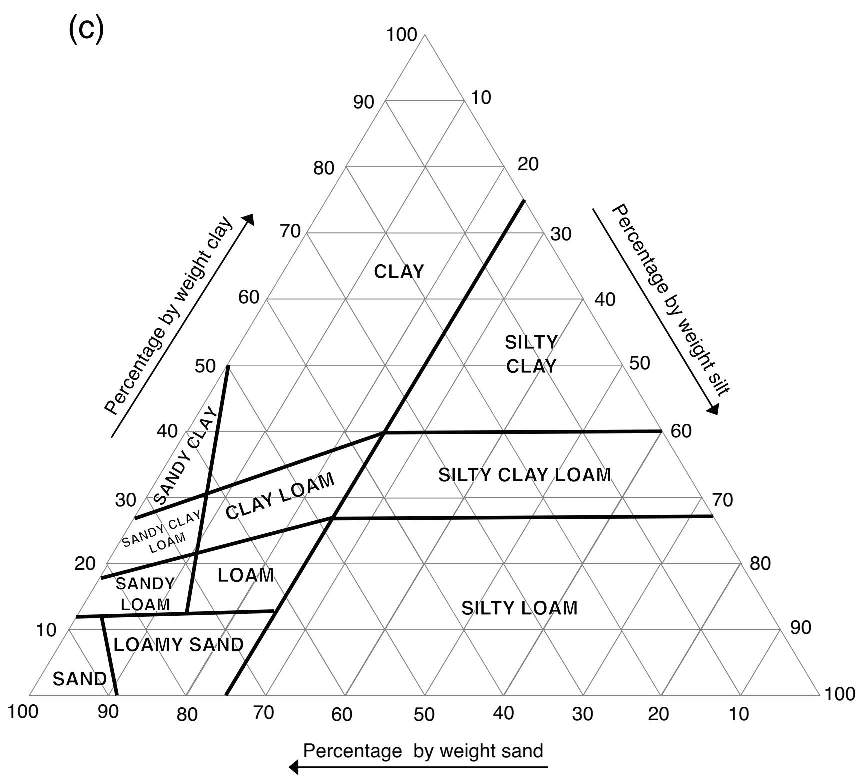


Figure 6.2 (Continued)

different classes depending on the relative proportion of sand, silt and clay. These vary by location (for example, the United Kingdom: Figure 6.2(a), and Australia: Figure 6.2(b)), but the principle is the same. The textural class of the soil is established by starting from a point on each axis that corresponds to the proportion represented by that grain size and then extending that line to the point where they intersect.

Vertical and horizontal heterogeneity in soil

Earlier we noted that the underlying geology is a key factor in determining the type of soil present at a location. We also mentioned that we should remember that soil arises from deposition of particles transported from further upstream in a catchment (by wind, water or ice) and over long time-scales. However, parent material is just one important factor, and there are several others at play in the process of soil formation (*pedogenesis*). These include topography, climate (especially rainfall and temperature), the action of organisms, and of course, time.

Let us now turn to considering **vertical heterogeneity** in soils. This means how different a soil is as you go down a profile. If you think back to when you may have seen a soil profile exposed, quite often you would have noticed distinct layering. Sometimes a darker, richer layer occurs at the top, with one or more layers of slightly different colour or form below that. Soil scientists call these **horizons**, each of which develops its own character (and therefore hydrological properties) over time. Soil scientists rely on the presence and characteristics of these horizons to classify and name soils beyond a simple textural class. Figure 6.3 shows the broad categories of horizons. The left figure shows vertical heterogeneity as a soil scientist might see it, whereas the right figure is a simplified conceptualisation and how the soil might be represented in a hydrological model. Where there is a deep litter layer, there is a high level of organic matter, sometimes sufficient to warrant its own horizon or 'O' horizon. Below this are the 'A' and 'B' horizons which may grade into 'C' (the **regolith** or consolidated rock). Figure 6.3 contains a clue as to why these layers are present anyway. Note the annotations which state 'zone of eluviation' and 'zone of illuviation'. Eluviation is the process whereby salts and other materials are leached out of the profile, while the zone of illuviation is where these are deposited or precipitated out.

These processes are directly related to the movement of water in the soil. The downwards movement of rainfall dissolves, mobilises and transports compounds in solution, which are then deposited lower down, thereby differentiating the soil horizons. Fluctuating moisture levels also impact upon the availability of air. When a soil is saturated, there is no air and therefore no oxygen. This means that the chemical environment turns from an oxidising to a reducing one, again changing the nature of the chemical processes occurring. Over time, these horizons develop as a result of the combination of the effects of fluctuating water (and oxygen) levels, but also as a result of other processes including those resulting from microbial and animal action.

The nature of soil layers is important to understand in hydrology because it influences that rate at which water passes through, or is stored within, the soil. For example, we know that soils with a high organic content in their upper layers are able to store more water than those with low organic content.

Now let us consider horizontal heterogeneity; how a soil varies spatially across the landscape. Even for soils with the same parent material, soil scientists have observed that the character of a soil varies markedly moving from the top of a hill downslope to the river valley bottom (Figure 6.4). This is known as a **catena**. At the top of the hill, soils might be well drained and reddish in colour. The reddish colour denotes oxidising conditions suggesting that the soil is typically aerated, which implies that it is well-drained. Conversely, soils towards the bottom of the valley might show blue/grey horizons indicating the reducing conditions

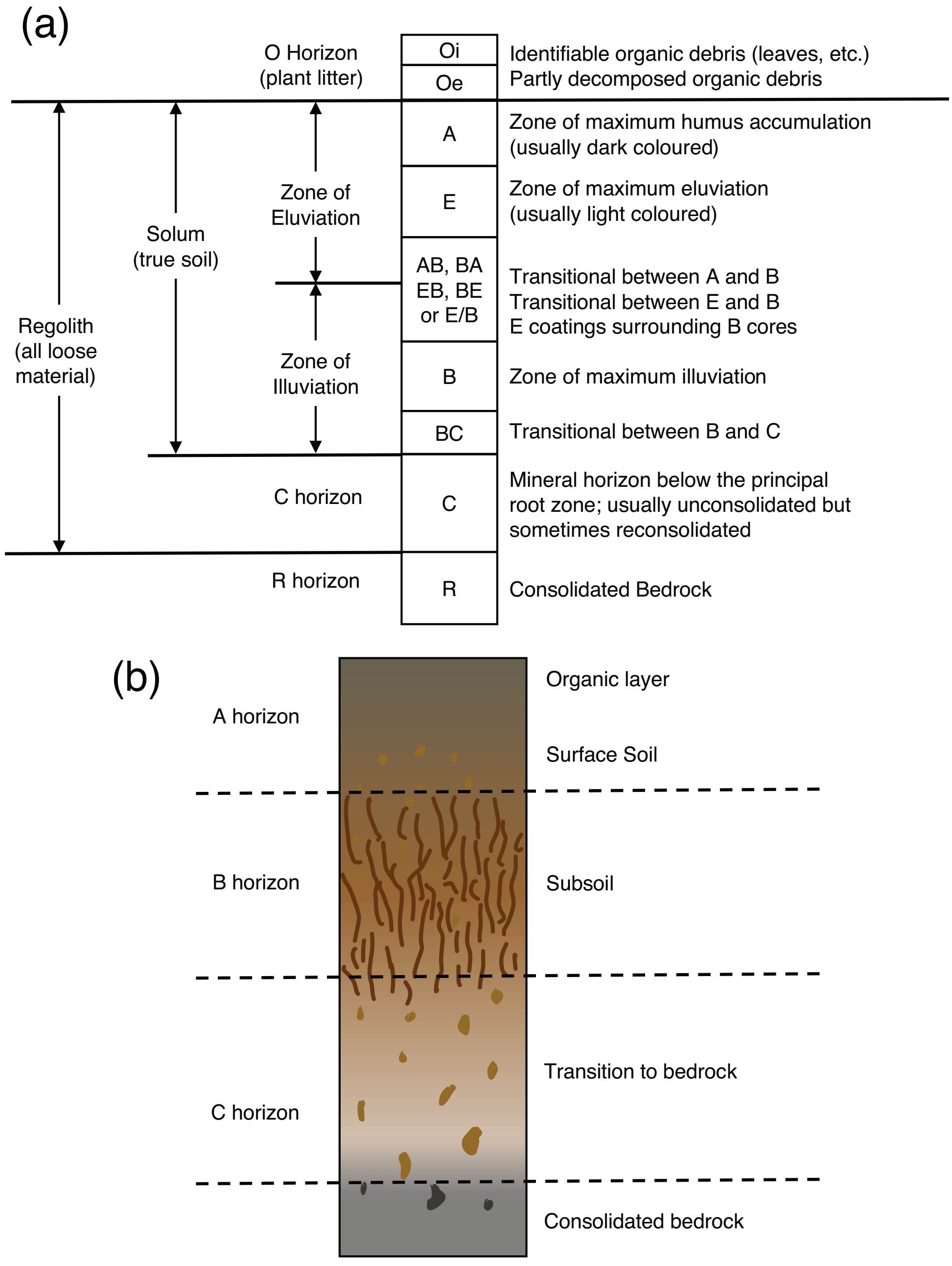


Figure 6.3 Vertical heterogeneity in soils showing soil horizons as represented by (a) a soil scientist and (b) how they might be represented more simply by a hydrologist in a hydrological model.

Source: Adapted from Ward and Trimble (2004)

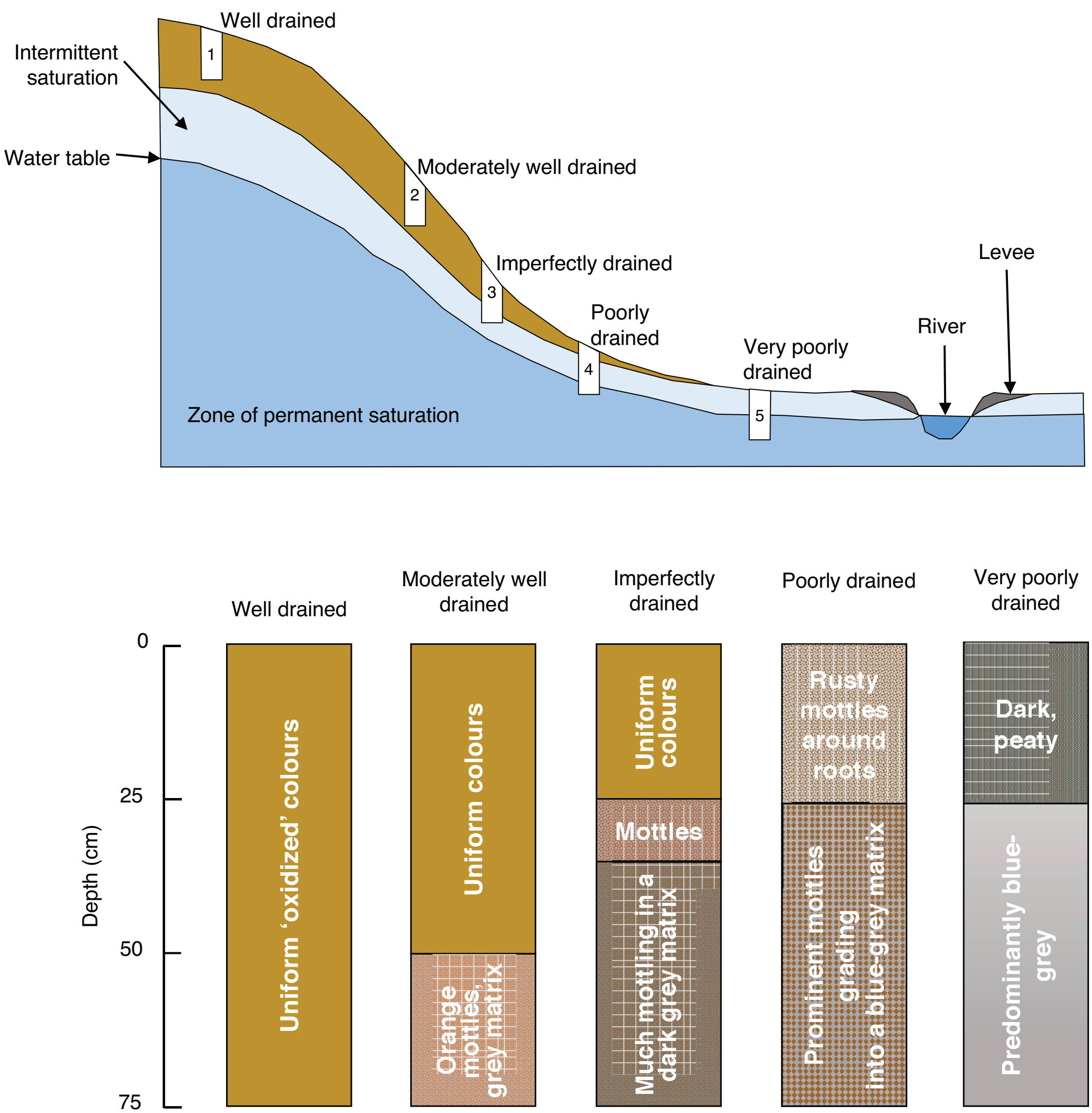


Figure 6.4 The concept of a catena showing the change of soils downslope in accordance with changes in water level and saturation.

Source: Reproduced from White (2006), with permission

which arise when the soil is waterlogged. Soils with intermediate levels of waterlogging are often darker with mottling and sometimes rusty coloured patches.

Figure 6.4 shows a catena in cross section, but when you imagine what soils mapped on this basis might look like in plan, you would expect groupings on the tops of hills and in river valleys, with

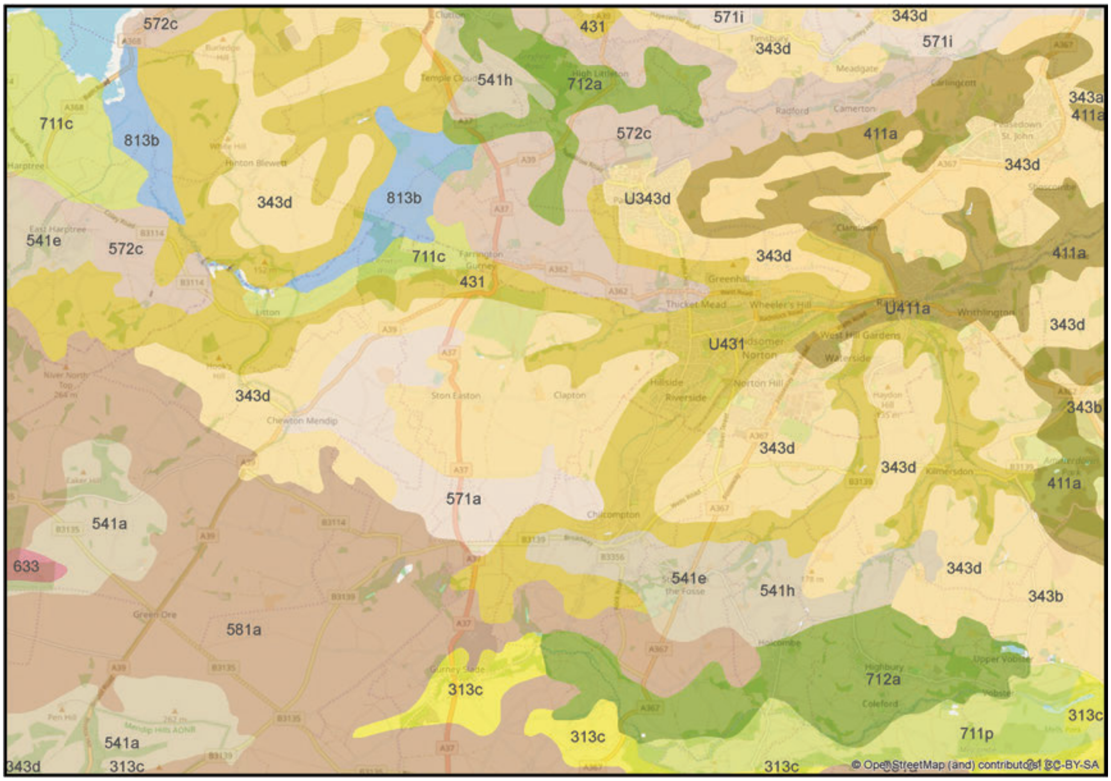


Figure 6.5 A section of a soils map for the United Kingdom illustrating the high spatial heterogeneity of soils (Soils data: Cranfield University, National Soil Resources Institute).

intermediate differentiation down slopes. Superimposing on this variation in topography a variation in underlying geology and climate, it is not difficult to appreciate why a soils map is highly heterogeneous (Figure 6.5). Given the importance of soils in regulating catchment response, obtaining a sufficiently representative understanding of the variation and characteristics of soils across a catchment is therefore of importance to any hydrological analysis of a catchment.

Soil structure

Earlier we introduced the idea of vertical heterogeneity in soils, indicating that various physical and chemical processes occur continuously. Not only can these processes give rise to soil horizons which are visibly different, they can also give rise to a distinct soil **structure**. By soil structure, we mean the orientation of soil particles or their grouping into larger formations, some of which have significant implications for the movement of moisture.

Movement of water in soils can occur in two domains. The first is the **macropore domain** – this refers to flow in cracks, cavities, animal burrows, old vegetation root pathways and the like (Figure 6.6). These structures are largely secondary and distinct from the fundamental structure of the soil. Flow can be quick in these structures and will tend to erode and extend these networks as it tries to find the easiest flow path. These pathways can also be fairly large – **soil pipes** are a good example of

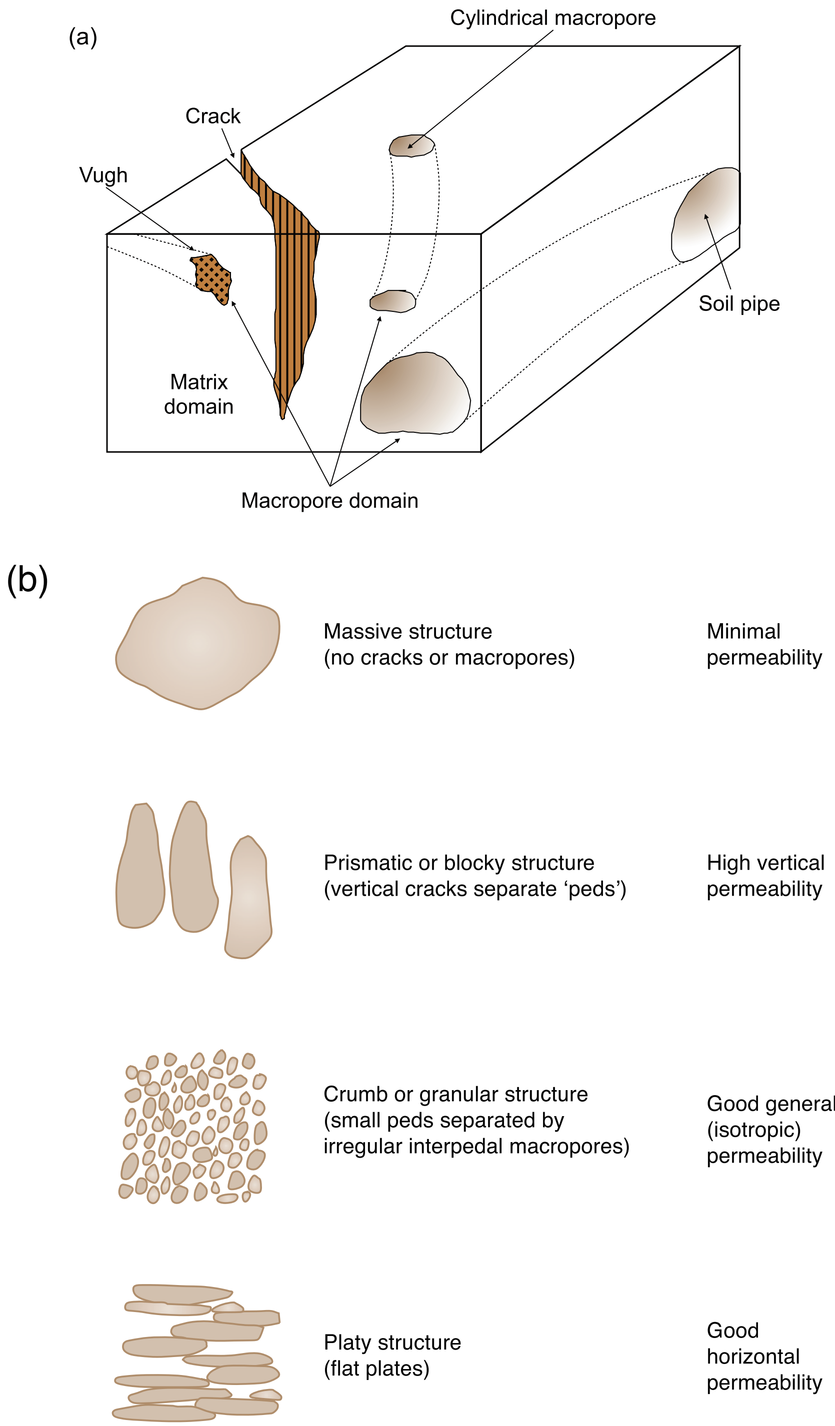


Figure 6.6 The macropore domain (a) characterised by voids of various shapes and origins, located within the matrix domain; (b) shows how a different structure in the matrix domain influences permeability.

Source: Redrawn from Jones (1997), with permission underground waterways which can quickly convey water and can be as large as 50 cm to 1 m in diameter in some locations. The extent to which flow in the macropore domain is significant in a catchment depends on the extent to which they are connected. The more connected they are, the quicker the likely hydrological response of the catchment (the time difference between rain seeping into the ground and an observable response in the flow in a channel). Because these structures are beneath the surface, it is very difficult to know what the likely contribution via macropore flow would be. These and other related challenges will be explored further in Chapter 7: Runoff.

The second domain is the **matrix domain**. By this we mean the dominant structure and bulk of the soil. We would describe a soil as **massive** if there is no discernible structure in any direction and no evidence of cracks or macropores. **Permeability** in all directions is therefore likely to be low. Permeability, in the general sense, means how easy it is for water to pass through a medium. In some soils, the B horizon forms into columnar or blocky structures. These prismatic shaped **peds** have vertical cracks separating them, providing a preferential pathway for flow. We would therefore expect high vertical permeability. In other soil types, for example neutral or slightly alkaline soils, the binding action of calcium and magnesium can result in soil particles grouping together as granules (Jones 1997). This granular or crumb-like structure leaves a network of spaces around the peds (**interpedal macropores**), which means that there is likely to be good permeability in all directions (**isotropic permeability**) (Jones 1997). In other types of soil, structure arises in the form of horizontally arranged plate-like structures. Here, horizontal permeability will obviously be high.

The key message from the discussion about the influence of soil structure on water movement is that we need to consider flow in both the matrix and macropore domains. These are of course not independent, but inter-related, flow pathways. When we simplify the complexity of soils by reducing it to a number of key parameters (as we will do in a later section), we are emphasising the matrix domain and ignoring the macropore domain. The reasons for this are those mentioned earlier; it is difficult to know the extent of macropore flow without detailed field investigation.

FUNDAMENTAL FORCES IN SOILS

In the previous chapter, we introduced the concept of **capillary forces**, referring to the example of water rising up a drinking straw (see p. 101, recap this before reading any further if you are unsure). You will recall that capillary forces are a consequence of the physics of small spaces. In small spaces such as soil and rock pores, **surface tension** (caused by **cohesion** within the liquid) and **adhesive** forces between the liquid and the solid particles act together to move water against gravity. Importantly these forces also act to hold water against gravity. As was the case with the straws of differing diameter, these forces are more pronounced the smaller the spaces. Since there is a relationship between the size of grains and the pores that are formed between them, this means that there is a general relationship between grain size and capillarity or the strength of this suction; the finer the grain size the stronger this force. In very fine grained soils such as clays, we would therefore expect to see greater amounts of water held against gravity. Up until now we have explained capillary action only in terms of surface tension and adhesion. There is another important concept in capillary action, **adsorption** (note it is a**d**sorption, not a**b**sorption which is something different). Adsorption is the force exerted through the electrostatic attraction between the faces of solid particles (which often carry a charge) and water molecules, which as we established in Chapter 1, are **dipolar**. The structure of many clay minerals is laminar – they are comprised of layers of silicates with gaps between them sufficient to allow water molecules to penetrate the structure of the mineral itself and become adsorbed to the internal surfaces (Figure 6.7). These internal surfaces can be significant – consider the differences in specific

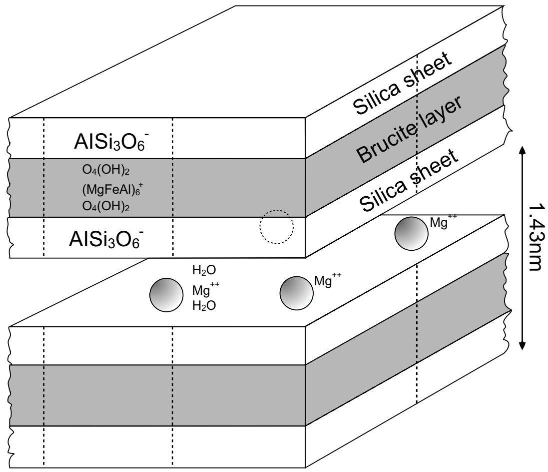


Figure 6.7 The structure of a common clay mineral (vermiculite) showing adsorption of water (see two water molecules attached to the Mg^{++}).

Source: Reproduced from White (2006), with permission

Table 6.1 Specific surface areas of particles and mineral types

Mineral or class size	*Specific surface (m^2/g)*
Coarse sand	0.01
Fine sand	0.1
Silt	1.0
Kaolinites	5–40
Hydrous micas	100–200
Iron and aluminium oxides	100–300
Vermiculites	300–500
Monmorillonite	750

Source: Simplified from White (2006)

surface (i.e. the total surface area per unit mass of material (m^2/g)) in Table 6.1. There are orders of magnitude differences between coarse sand, fine sand and silt, but many types of clays have surface areas several hundred times that of silt. The important point here is that water is able to penetrate the structure of many clays and there is a significant internal structure for adsorption to occur on. Some clays will swell and expand as a result. In addition to the capillary action arising from adsorption there is also the effect of surface tension and adhesion. As a result, clay can hold significant quantities of water but adsorptive forces mean the water does not pass through, or leave, the clay easily. Bear in mind these fundamental forces as we consider representing soils in the next section by way of a few key parameters.

Representing soil properties through key numerical concepts

Porosity

You will recall in the previous chapter we introduced the concept of **porosity**, which applies to both rock and soil. For convenience, we'll repeat the basic definition here. This is defined simply as the proportion of a soil or rock that consists of open spaces, and must therefore logically be the volume of the pores divided by the total volume of the rock or soil, often represented using the symbol 'n' (Equation 6.1).

$$n = \frac{V_p}{V_t} \tag{6.1}$$

Where n is porosity of a sample (dimensionless or e.g. cm^3/cm^3), V_p is the volume of pores in the sample (e.g. cm^3) and V_t is the total volume of the sample (e.g. cm^3). You will recall that porosity is usually expressed as a decimal fraction (between 0 and 1), but it can also be expressed as a percentage (see Table 5.1 in the previous chapter for an indication of the typical range in porosity of different types of rock and soil).

Saturation

When the volume of pores is completely filled with water, the soil is said to be saturated and this obviously also defines the maximum amount of water that can be in a soil (*saturated water content*). One way of expressing soil water content is as a percentage of the saturated content and is a useful method of telling how wet the soil actually is.

Volumetric soil water content

Soil water content is normally expressed as a **volumetric soil moisture content** or soil moisture fraction and given the Greek symbol theta (θ) – Equation 6.2.

$$\theta = \frac{V_w}{V_t} \tag{6.2}$$

where V_w is the volume of water in a soil sample and V_t is the total volume of the soil sample.

This is normally written as a decimal fraction (i.e. 0.4 = 40 per cent water by volume). As θ is a volume divided by a volume it has no units, although it is sometimes denoted as m^3/m^3. If you know the depth of the soil, the volumetric water content of a soil can be converted into an equivalent depth of water and therefore easily related to other equivalent depths such as rainfall and evaporation. For example, if θ = 0.4 and the soil depth is 500 mm, there is an equivalent depth of water of 200 mm (0.4 × 500 mm = 200 mm). Thinking about the equivalent depth of water in the soil in this way is a fundamental approach in soil moisture budgeting and some types of hydrological modelling.

Field capacity

In theory, water can fill all of the pores in a soil; therefore porosity is the maximum potential volumetric water content. In practice the volumetric soil moisture seldom reaches the porosity value and if it does, gravity acts on the water to force drainage through the profile that quickly drops moisture levels back below porosity. **Field capacity** is the stable point of saturation after rapid drainage, normally a day to a couple of days. In other words, field capacity is the amount of water held back in the soil against gravity drainage by capillary forces, and for this reason, another term to describe this moisture state is the **drained upper limit**. Like porosity, field capacity is also expressed as a decimal fraction between 0 and 1. As indicated above, it is the strength of capillary forces that determines the amount of water at field capacity (usually denoted by θ_{fc}), and consequently you would expect higher values of field capacity for fine grained materials such as clays. As was the case for calculating the amount of water at saturation in the example above, calculating the amount of water in a soil of known depth at field capacity is just as straightforward. If θ_{fc} = 0.55 and the soil depth is 400 mm, there is an equivalent depth of water of 220 mm (0.55 × 400 mm = 220 mm).

Wilting point

Wilting point is a term derived from agriculture and refers to the soil water content when plants start to die back (wilt). This is significant in hydrological processes as beyond this point the plants will no longer transpire. Some plants can recover from this point if water is provided, although for most plants experiencing moisture conditions towards this end of the soil moisture gradient invariably means that plant stress is occurring. If this is an agricultural crop there would be likely impacts in terms of crop yield and crop quality. Like porosity and field capacity, θ_{wp} is also a decimal fraction. Although there is moisture in the soil below the wilting point, this moisture is held so tightly by the soil it is not accessible to plants, and would require oven drying to remove it. Because these values are on a moisture gradient, the following must hold true:

$$n > \theta_{fc} > \theta_{wp}$$

Table 6.2 provides typical values of porosity, field capacity and wilting point for a range of soil textural classes.

Plant available water

Plant available water is simply the volume of water that plants have access to. This is also known as *available water capacity*. As indicated above, water

Table 6.2 Key soil parameters for a range of soil textural classes

Soil texture class	*% silt*	*% sand*	*% clay*	k_{sat} *(mm/hr)*	*n*	θ_{fc}	θ_{wp}
Sand	5	92	3	88.1	0.373	0.150	0.033
Loamy sand	12	82	6	63.0	0.386	0.191	0.051
Sandy loam	32	58	10	29.6	0.419	0.277	0.091
Loam	39	43	18	17.0	0.437	0.341	0.144
Silty loam	70	17	13	8.6	0.476	0.439	0.164
Sandy clay loam	15	58	27	23.1	0.413	0.307	0.153
Clay loam	34	32	34	9.8	0.447	0.384	0.211
Silty clay loam	56	10	34	5.2	0.478	0.452	0.249
Sandy clay	6	52	42	15.5	0.416	0.333	0.197
Silty clay	47	6	47	3.8	0.479	0.456	0.283
Clay	20	22	58	5.1	0.452	0.408	0.270

Note: Percentage composition taken as the centroid of the idealised textural class.

Source: http://biocycle.atmos.colostate.edu/shiny/soils. This is an interactive app which allows the user to explore soil parameters dynamically

at contents below wilting point is not available to plants, because it is bound too tightly to the soil. In contrast, water above field capacity generally drains away rapidly, so in effect is not really available to plants. Logically then, the water that is available for transpiration is Equation 6.3:

$$\theta_{paw} = \theta_{fc} - \theta_{wp} \tag{6.3}$$

So, in a clay-loam soil that is 600 mm deep, with a field capacity of 0.35 and a wilting point of 0.15, what is the plant available water? Remember that all these soil parameters represent proportions by volume, so if you know the depth of the soil you can easily turn this into an equivalent depth of water (θ_{paw} = (600 mm × 0.35) – (600 mm × 0.15) = 120 mm).

Figure 6.8 represents a more graphical view of data similar to that presented in Table 6.2. It can be seen that while a clay soil has a high porosity, it also has a high wilting point, which means low plant available water. This is because plants struggle to overcome the strong adhesive forces exerted by clay particles on the water held between them. Similarly, at the opposite end of the graphic at the higher grain sizes, sand has a lower wilting point, but also a lower porosity, therefore also leading to low plant available water. Soils with the highest plant available water occur in the intermediate grain sizes, and these correspond to the loams preferred for horticulture.

Soil moisture deficit

Soil moisture deficit is the amount of water required (in mm depth) to fill the soil up to field capacity. This is an important concept in both agriculture and hydrology. Efficient irrigation, for example, should provide enough water to take the soil moisture to field capacity. Any irrigation beyond this point leads to saturation (which may not be desirable for certain crops anyway), but more importantly will drain away and not be available to plants, representing 'wasted' water from an agricultural perspective and may contain soluble nutrients that can create problems from a water quality perspective. This is also an important hydrological parameter as it is often assumed that all rainfall infiltrates into a soil until the moisture content reaches field capacity. The soil moisture deficit

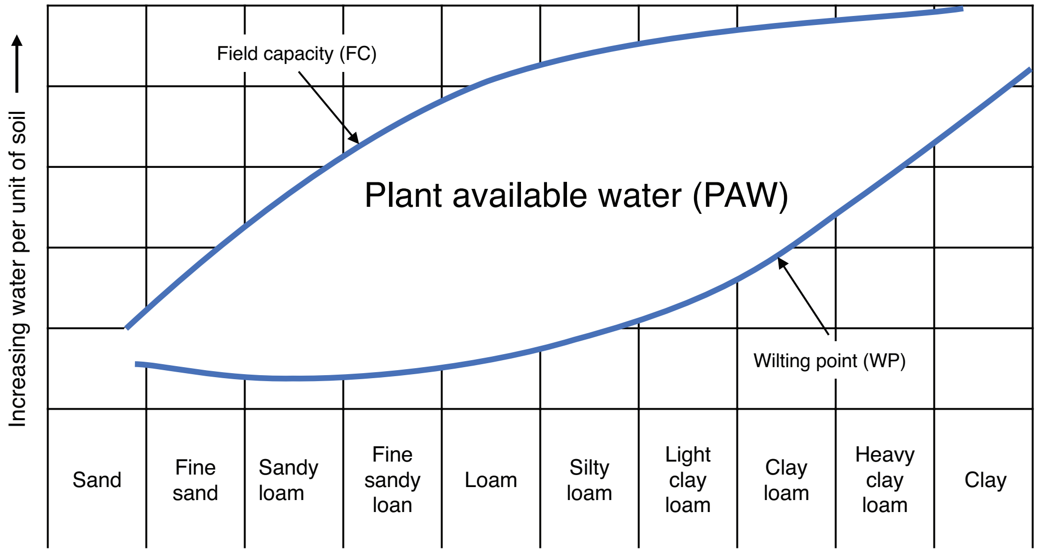


Figure 6.8 The textural class of the soil determines the porosity, field capacity and wilting point of the soil, which in turn determines the plant available water. Highest plant available water occurs in intermediate grain sizes.

Source: Adapted from Ward and Trimble (2004)

gives an indication of how much rain is required before saturation, and therefore when overland flow may occur (see Chapter 7). Soil moisture deficit can be calculated by the following equation (Equation 6.4), where θ is defined as the current moisture condition:

$$\theta_{smd} = \theta_{fc} - \theta \qquad (6.4)$$

Soil moisture tension

Earlier we explained why capillary action holds back water against gravity in soils (see 'Fundamental forces in soils'). When this force acts over an area, it is defined as pressure (i.e. pressure is force per unit area). Capillary forces therefore result in suction (negative pressure), which is also referred to as **soil moisture tension**. This property of a soil can be measured directly (see later section on measurement). The symbol ψ (psi), is used to represent this negative pressure head and is also called **matric suction**, **tension head** or **matric potential**.

Figure 6.9 represents the various moisture condition parameters (saturation, field capacity, wilting point) using the analogy of a glass filled with water and ice cubes – where the ice cubes are the soil particles. The strength of soil suction at any point in time depends firstly on the pore size distribution (i.e. determined by the textural properties of the soil) and secondly on the amount of water present in the soil. The annotation on the right of Figure 6.9 shows that when the soil is saturated, $\psi = 0$. Water between saturation and field capacity is not held by capillary forces and is free to move (drain). For this reason, it is known as **gravitational water**. As discussed previously, water between the moisture conditions of field capacity and wilting point is held by the soil, but is available for transpiration and is therefore **plant available water**. Notice though, that the negative pressure head (ψ) increases sharply from typically –340 cm of water at field capacity to –15,000 cm of water at field capacity. The latter value is the suction equivalent of the pressure exerted by a home espresso machine (15 bars). In other words, at wilting point, moisture is being held back by the soil by a suction of the same order of magnitude of pressure that water is passed through coffee to make espresso.

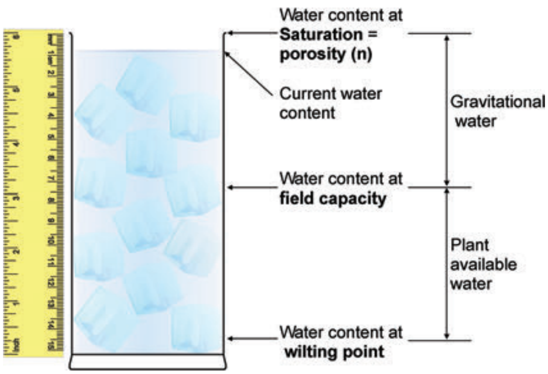


Figure 6.9 Concepts of soil moisture illustrated using the analogy of a glass filled with ice and water, where the ice represents the soil particles. The ruler represents the fact that if something is known about the textural properties and depth of the soil, then the equivalent depth of water can be calculated (mm). On the right of the figure the effect of decreasing moisture condition on soil tension is shown, with reducing moisture conditions resulting in significantly increasing soil moisture tension – water becomes increasingly tightly bound to the soil.

So it is clear that reducing moisture conditions in the soil means that the water that is held there, is held by an ever-increasing force. In other words, this is a non-linear relationship. Earlier we said that this is also a function of the pore size distribution. This means that every type of soil will have its own characteristic relationship between moisture content and **tension head** or **matric potential**. These are called **soil moisture curves** (or **suction–moisture curves**) and are typically generalised per soil textural class. Figure 6.10 shows soil moisture curves for a clay and sand, representing the range in behaviours across soil textures.

Although the soil moisture curves in Figure 6.10 are represented by a single line, it is a little more complicated than that. The shape of this curve actually depends on whether you are measuring soil tension as you are wetting the soil, or as you are drying it. This phenomenon is known as **hysteresis**, and is related to the way that water enters and leaves pores. It takes a larger force for air to exit a narrow pore neck (e.g. when it is drying out) than for water to enter (wetting) (Figure 6.11). Care must therefore be taken in interpreting a suction–moisture curve, as the method of measurement may have a large influence on the overall shape.

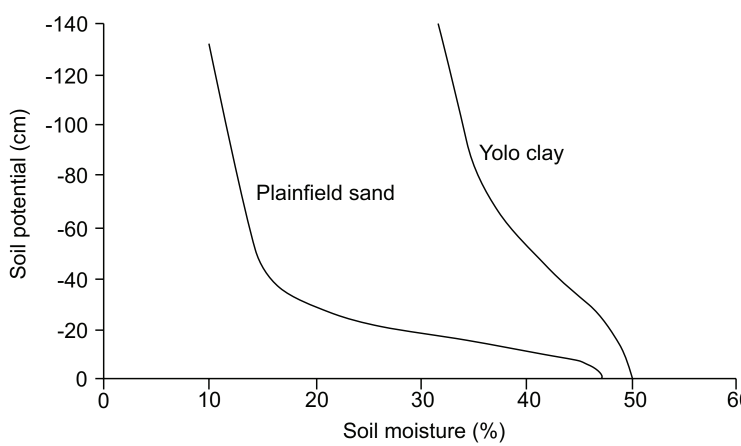


Figure 6.10 Soil moisture characteristic (matric suction) curves for different soil textures. Note the non-linear nature of the curves.

Source: Redrawn from Thompson (1999), with permission

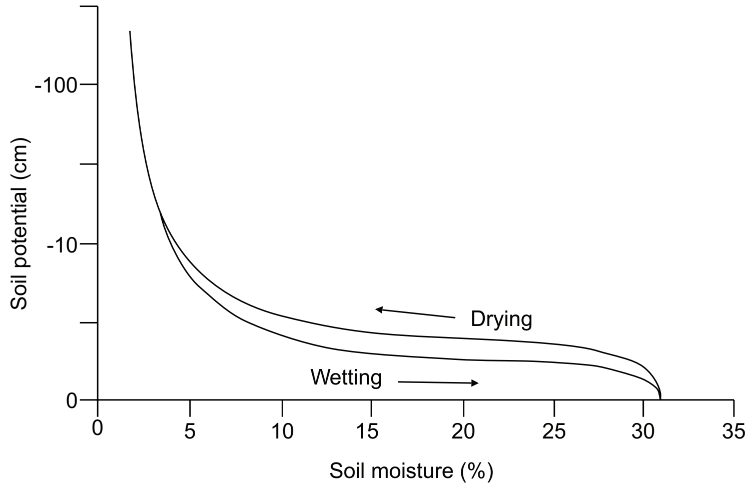


Figure 6.11 A generalised suction–moisture (or soil characteristic) curve for a soil. The two lines show the difference in measurements obtained through a wetting or drying measurement route (hysteresis).

Source: Redrawn from Thompson (1999), with permission

How does this influence movement of water in soil?

In the previous chapter (groundwater), you were introduced to the concept of hydraulic head and Bernoulli's equation. For movement in unsaturated soils, Equation 5.3 in the previous chapter (which was for saturated conditions), can be re-written as Equation 6.5:

$$h = z + \psi \tag{6.5}$$

where h is the total potential (head), z is the gravitational potential and ψ is the matric potential.

Water movement in soils therefore occurs as both a function of gravity and matric potential. At saturation, matric potential is zero, so flow occurs as saturated flow and according to Darcy's Law. As soon as the soil is unsaturated, matric potential starts to counter the force of gravity, increasingly so as the soil dries. The implication of this is that hydraulic conductivity in unsaturated soils (K_{unsat}) must vary, unlike saturated conditions in groundwater, where it is constant for a given material (K_{sat}). As can be seen in Figure 6.12, unsaturated hydraulic conductivity changes in a non-linear way depending on the

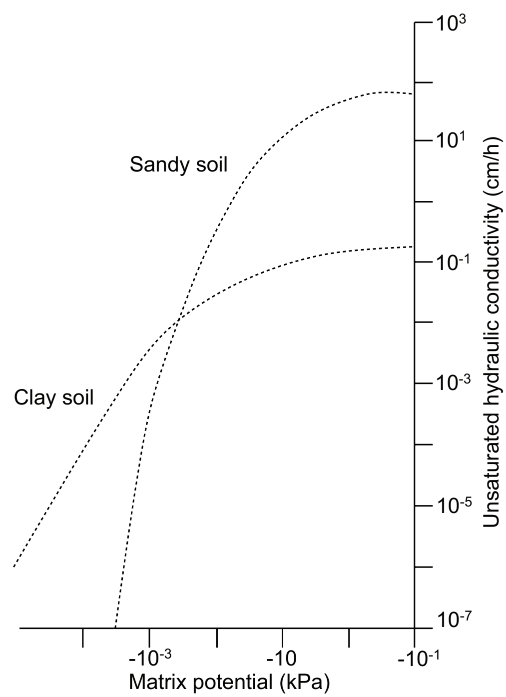


Figure 6.12 Soil moisture tension and unsaturated hydraulic conductivity.

Source: Reproduced from White (2006), with permission prevailing soil tension, which in turn depends on the soil moisture at any one time. Because hydraulic conductivity is a function of soil moisture which changes constantly in the soil, modelling movement of water in the unsaturated zone is much more complicated than for the saturated zone (groundwater). As the soil dries, hydraulic conductivity reduces in a non-linear way and water movement slows considerably. All other things being equal, movement of water under unsaturated conditions is much slower than saturated conditions. Conversely, small increases in water content can result in significantly faster movement. This phenomenon also makes things numerically complex, and so Richards' approximation of Darcy's equation is often used in unsaturated conditions. These additional complications are the reason we introduced you to the principles of groundwater flow in the previous chapter first.

This brings us to the end of the section dealing with the fundamental forces and principles. We will now apply these principles in considering firstly how water enters the soil (**infiltration**) and secondly how it is then redistributed in the soil profile (**percolation** or **soil moisture redistribution**), some of which will enter the saturated zone and therefore become groundwater via **recharge**.

Infiltration and soil moisture redistribution

Infiltration is the process whereby rainfall or ponded water penetrates the surface of the soil and become soil moisture. How much water enters a soil during a certain time interval is known as the **infiltration rate**, and is dependent on the current water content of the soil and the ability of a soil to transmit the water. Although it is a property of the soil itself, it also depends importantly on the nature of the surface covering and is strongly influenced by the history of land management at the site. The rate at which this occurs in a rainfall event is not constant. Generally, water initially infiltrates at a faster rate and slows down with time (see Figure 6.13). When the infiltration rate slows down to a steady level (where the curve flattens off in Figure 6.13) the **infiltration capacity** has been reached. This is the rate of infiltration when the soil is fully saturated. The terminology of infiltration capacity is misleading as it suggests a capacity value rather than a rate. In fact, infiltration capacity is the infiltration rate when the soil is filled to capacity with water. The curves shown in Figure 6.13 are sometimes called the Philip curves, after Philip (1957) who built upon the pioneering work of Horton (1933) and provided sound theory for the infiltration of water.

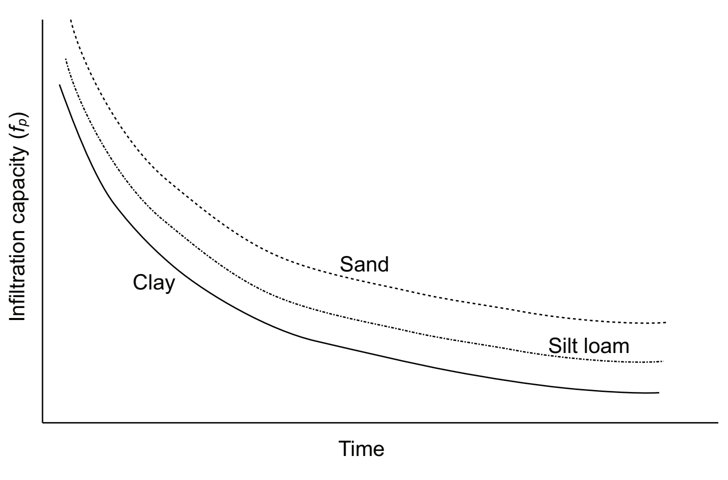


Figure 6.13 Generalised infiltration curves for a sand and a clay soil.

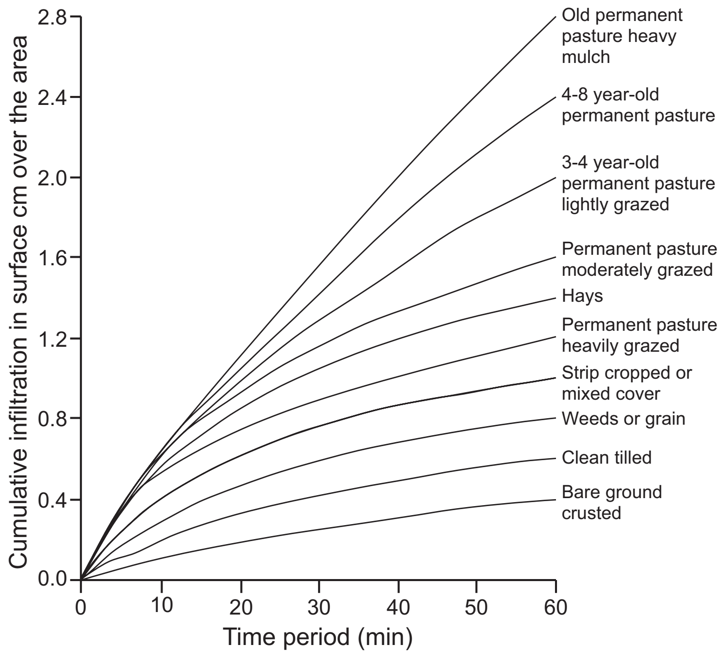


Figure 6.14 The influence of land use and land cover on infiltration rates.
Source: Reproduced from Ward and Trimble (2004), with permission

Infiltration capacity is sometimes referred to as the saturated hydraulic conductivity. This is not absolutely true as the measurement is dependent on the amount of water that may be ponded on the surface creating a high hydraulic head. Saturated hydraulic conductivity should be independent of this ponded head of water. There are conditions when infiltration capacity equals saturated hydraulic conductivity, but this is not always the case.

Infiltration is strongly related to land use and land-cover. Figure 6.14, for example, shows that infiltration through old permanent pasture with heavy mulch (layer or organic matter) could have as much as seven times the infiltration after an hour in comparison with bare crusted ground.

Other factors, such as **antecedent moisture condition** are also important. The antecedent moisture condition refers to how wet the soil already was at the start of a rainfall event. A drier soil will have faster infiltration than a wetter soil. Biological action can be important in helping maintain an open soil structure, as can the presence of organic matter which improves infiltration. Over-stocking and compaction by vehicles in contrast can considerably reduce infiltration, which means that rather than replenishing soil moisture, rainfall just becomes surface runoff. Some types of clay soils are prone to cracking and crusting and others form surface seals, quite often due to the clay minerals present.

The main force driving infiltration is gravity, but it may not be the only force. When soil is very dry it exerts suction that will draw the infiltrating water towards the drier area. Consider rain falling into the surface soil crack in Figure 6.15 – gravitational forces will draw the water downwards, but capillary forces will also be drawing the water laterally. As the moisture penetrates further, gravitational forces and capillary forces will be acting in the same direction and drawing water downwards. With both of

Figure 6.15 A neutron probe sitting on an access tube. The black cable extends down into the tube with the source of fast neutrons (and counter) at the tip.

these forces acting in concert, the infiltrating water moves down through the soil profile in a wetting front. The wetting front is three-dimensional, as the water moves outwards as well as vertically down. The shape of the curve in Figure 6.13 is related to the speed at which the wetting front is moving. It slows down the further it gets away from the surface as it takes longer for the water at the surface to feed the front (and as the front increases in size).

The ability of a soil to transmit water is dependent on the pore sizes within it and most importantly on the connections between pores. Pores can be classified according to size or function (McLaren and Cameron 1996). Macropores are defined as pores greater than 30 μm (microns) in diameter but can also be defined by their drainage characteristic (the amount of pressure required to remove water from the pore). A well-structured soil consists of stable aggregates with a wide range of pore sizes within and between the aggregates. In this case macropores may make up at least 10 per cent of this soil volume. This structure provides numerous interconnected pathways for the flow of water with a wide range of velocities. In less well-structured soils, biological activity (e.g. roots and worms) can produce macropores that provide flowpaths for water that are largely separated from the main soil matrix (Clothier et al. 1998). These are essentially two different types of macropores: those large pores within the soil matrix (**matrix domain**); and those that are essentially separated from the matrix (**macropore domain**).

In flat areas of the landscape, the direction of water movement will obviously be vertically downwards. On slopes there will be elements of vertical and downslope movement, the net movement being the vector of the vertical and slope characteristics. Flow downslope is called **lateral flow**, **transverse flow** or **interflow**.

When surface soil is drier than deeper soil, water movement can be in the opposite direction against gravity. You will recall this also occurs at the water table, where the **capillary fringe** defines the transitional area between the phreatic and vadose zones. Where soil water moves back onto the surface from the soil zone, this is known as **exfiltration** and as we will see in the next chapter, can be important in surface runoff.

MEASUREMENT OF WATER IN THE SOIL

Measuring soil moisture

Although the earlier section referred to **volumetric water content**, another way of representing soil water content is in relation to its mass. This is known as **gravimetric soil moisture content** (G). Gravimetric soil moisture content is the ratio of the weight of water in a soil to the overall weight of the soil (Equation 6.6):

$$G = \frac{M_w - M_d}{M_d} = \frac{\theta}{\rho_b \rho_w} \tag{6.6}$$

where G is the gravimetric water content (g/g), M_w is the mass of the wet sample, and M_d is the mass of the dry sample. The second part of the equation shows how gravimetric and volumetric soil water content can be related to each other by the soil bulk density: the density of soil in situ and where ρ_b is the bulk density of soil (g/cm^3); and ρ_w is the density of water (g/cm^3). As the density of water is close to 1 g/cm^3 it can be ignored. Soil bulk density (ρ_b) is the ratio of the mass of dry soil to the total volume of the soil (Equation 6.7).

$$\rho_b = \frac{M_d}{V_t} \tag{6.7}$$

As described above, the density of water is very close to 1 g/cm^3 (but temperature dependent; see Figure 1.3), therefore the weight of water is often assumed to be the same as the volume of water. The same cannot be said for soil: the bulk density depends on the mineralogy and packing of particles so that the volume does not equal the weight. Soil bulk density gives an indication of soil compaction with a cultivated topsoil having a value of around 1 g/cm^3 and a compacted subsoil being as high as 1.6 g/cm^3 (McLaren and Cameron 1996). It is important to note that because of this, gravimetric soil moisture content is not the same as volumetric soil moisture content, and care must be taken in distinguishing between them as they are not interchangeable terms. These concepts underpin the gravimetric method outlined below.

Gravimetric method

The simplest and most accurate means for the measurement of soil water is using the gravimetric method. This involves taking a soil sample, weighing it wet, drying in an oven and then weighing it dry. Standard practice for the drying of soils is 24 hours at 105 °C (Gardner 1986). The difference between the wet and dry weights tells you how wet the soil was. If it is volumetric soil moisture content that is required then you must take a sample of known volume. This is commonly done using an undisturbed soil sample and utilising Equation 6.8.

$$\theta = \frac{V_w}{V_t} \approx \frac{M_w - M_d}{V_t} \tag{6.8}$$

Where θ is the volumetric water content (cm^3/cm^3); V_w is the volume of water (cm^3); V_t is the total volume of soil (cm^3); M_w is the mass of water (g); and M_d is the mass of dry soil (g). The density of water is close to 1 g/cm^3 so the weight of water can be assumed to equal the volume of water; hence the ≈ symbol in the equation above.

Gravimetric analysis is simple and accurate but does have several drawbacks. Most notable of these is that it is a destructive sampling method and therefore it cannot be repeated on the same soil sample. This may be a problem where there is a requirement for long-term monitoring of soil moisture. In this case a non-destructive moisture-sampling method is required. There are three methods that fit this bill, but they are indirect estimates of soil moisture rather than direct measurements as they rely on measuring other properties of soil in water. The three methods are: **neutron probes**, *electrical resistance blocks* and **time domain reflectometry**. All of these can give good results for monitoring soil moisture content, but are indirect and require calibration against the gravimetric technique.

Neutron probe

A neutron probe has a radioactive source that is lowered into an augured hole; normally the hole is kept in place as a permanent access tube using aluminium tubing. The radioactive source emits fast (or high energy) neutrons that collide with soil and water particles. The fast neutrons are very similar in size to a hydrogen ion (H^+ formed in the disassociation of the water molecule) so that when they collide the fast neutron slows down and the hydrogen ion speeds up. In contrast, when a fast neutron collides with a much larger soil particle it bounces off with very little loss of momentum. The analogy can be drawn to a pool table. When the cue ball (i.e. a fast neutron) collides with a coloured pool ball (i.e. a water particle) they both move off at similar speeds, the cue ball has slowed down and the coloured ball has speeded up. In contrast, if the cue ball hits the cushion on the edge of a pool table (i.e. a soil particle) it bounces off with very little loss of speed. Consequently the more water there is in a soil the more fast neutrons would slow down to become 'slow neutrons'. A neutron probe counts the number of slow neutrons returning towards the radioactive tip, and this can be related to the soil moisture content. The neutron probe readings need to be calibrated against samples of soil with known moisture contents. This is often done by using gravimetric analysis on the samples collected while the access tubes are being put in place. It is also possible to calibrate the probe using reconstituted soil in a drum or similar vessel. It is important that the calibration occurs on the soil actually being measured, since the fast:slow neutron ratio will vary according to mineralogy of the soil.

Although the neutron probe is essentially non-destructive in its measurement of soil water content, it is not continuous. There is a requirement for an operator to spend time in the field taking measurements at set intervals. This may present difficulties in the long-term monitoring of soil moisture. Another difficulty with a neutron probe is that the neutrons emitted from the radioactive tip move outwards in a spherical shape. When the probe tip is near the surface some of the sphere of neutrons will leave the soil and enter the atmosphere, distorting the reading of returning slow neutrons. A very careful calibration has to take place for near-surface readings and caution must be exercised when interpreting these results. This is unfortunate as it is often the near-surface soil moisture content that is of greatest importance. Although neutron probes are reliable instruments for the monitoring of soil moisture, the cost of the instruments, difficulties over installing access tubes (Figure 6.15), calibration problems and the near-surface problem have meant that they have seldom been used outside a research environment.

Electrical resistance blocks

Electrical resistance blocks use a measurement of electrical resistance to infer the water content of a soil. As water is a conductor of electricity it is reasonable to assume that the more water there is in a soil the lower the electrical resistance, or conversely, the higher the electrical conductivity. For this instrumentation two small blocks of gypsum are inserted into the soil and a continuous measurement of electrical resistance between the blocks is recorded. The measure of electrical resistance can be calibrated against gravimetric analysis of soil moisture. The continuity of measurement in electrical resistance blocks is a great advantage of the method, but there are several problems in interpreting the data. The main difficulty is that the conductivity of the water is dependent on the amount of dissolved ions contained within it. If this varies, say through the application of fertiliser, then the electrical resistance will decrease in a manner unrelated to the amount of water present. The second major difficulty is that the gypsum blocks deteriorate with time so that their electrical conductivity alters. This makes for a gradually changing signal, requiring constant recalibration. The ideal situation for the use of electrical resistance blocks is where they do not sit in wet soil for long periods and the water moving through the soil is of relatively constant dissolved solids load. An example of this type of situation is in sand dunes, but these are not particularly representative of general land use.

Time domain reflectometry

Time domain reflectometry (TDR) is a more recent soil moisture measurement technique. The principle of measurement is that as a wave of electromagnetic energy is passed through a soil the wave properties will alter. The way that these wave properties alter will vary, dependent on the water content of the soil. TDR measures the properties of microwaves as they are passed through a soil and relates this to the soil moisture content. Although this sounds relatively simple it is a complicated technique that requires detailed electronic technology. Up until the late 1990s this had restricted the usage of TDR to laboratory experiments but there are now soil moisture probes available that are small, robust and reliable in a field situation. An example of this is the Theta probe shown in Figure 6.16.

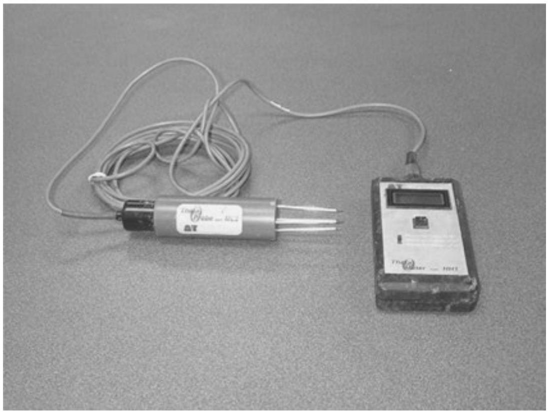


Figure 6.16 The Theta probe (manufactured by Delta-T devices). An example of a small, time domain reflectometry instrument used to measure soil moisture content in the field. The metal spikes are pushed into the soil and the moisture level surrounding them is measured.

Measuring soil suction

A **tensiometer** is used to measure the soil suction pressure or soil moisture tension. This is the force exerted by capillary forces and it increases as the soil dries out. A tensiometer is a small ceramic cup on the end of a sealed tube of water. The dry soil attempts to suck the water from the water-filled tube through the ceramic cup. At the top of the tube a diaphragm measures the pressure exerted by this suction. Tensiometers also have the ability to measure a positive pressure when the soil water is held under pressure (e.g. during a rising water table). The units of soil suction are Pascals, the SI units for pressure.

A **suction–moisture curve** (see Figure 6.11) can also be derived for a soil sample using a pressure plate apparatus. This uses a pressure chamber to increase the air pressure surrounding a soil sample and force water out of pores and through a ceramic plate at its base. When no more water can be forced out then it is assumed that the capillary forces (i.e. the soil suction) equal the air pressure and the sample can be weighed to measure the moisture content. By steadily increasing the air pressure between soil moisture measurements, a suction–moisture curve can be derived. This can be interpreted to give important information on soil pore sizes and is also important for deriving an unsaturated hydraulic conductivity value for a given soil moisture (Klute 1986).

Measuring infiltration

Infiltration rate is measured by recording the rate at which water enters the soil. There are numerous methods available to do this, the simplest being a ring **infiltrometer** (see Figure 6.17). A solid ring is pushed into the ground and a pond of water sits on the soil (within the ring). This pond of water is kept at a steady level by a reservoir held above the ring. Recordings of the level of water in the reservoir (with time) give a record of the infiltration rate. To turn the infiltration volume into an infiltration depth the volume of water needs to be divided by the cross-sectional area of the ring.

A simple ring infiltrometer provides a measure of the ponded infiltration rate, but there are several associated problems. The first is that by using a single ring a large amount of water may escape around the sides of the ring, giving higher readings than would be obtained from a completely saturated surface. To overcome this a double ring infiltrometer is sometimes used. With this, a second wider ring is placed around the first and filled with water so that the area surrounding the measured ring is continually wet. The second problem is that ponded infiltration is a relatively rare event across a catchment. It is more common for rainfall to infiltrate directly without causing a pond to form on the surface. To overcome this a rainfall simulator may be used to provide the infiltrating water.

Estimating soil water

In the previous section it was stated that several of the methods listed were indirect measurement methods or estimation techniques. They certainly do not measure soil moisture content directly, but

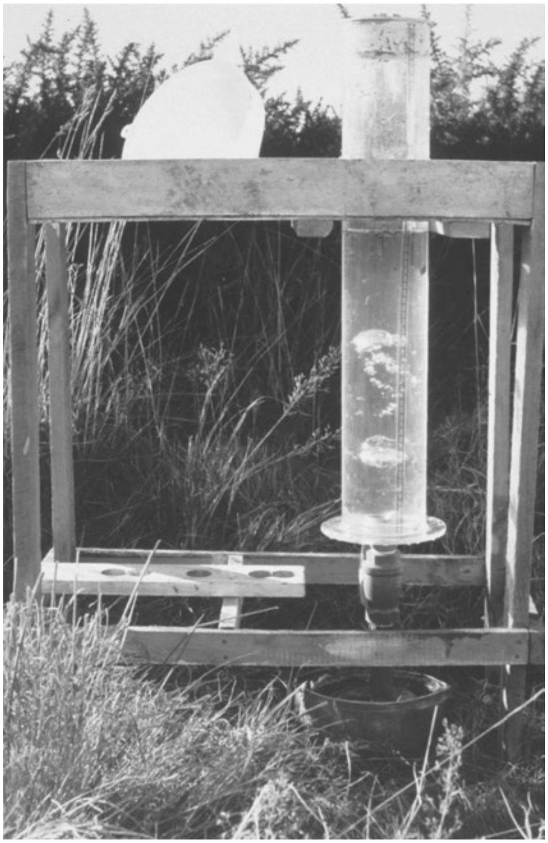


Figure 6.17 A single ring infiltrometer. The ring has been placed on the ground and a pond of water is maintained in the ring by the reservoir above. A bubble of air is moving up the reservoir as the water level in the pond has dropped below the bottom of the reservoir. A reading of water volume in the reservoir is taken and the time recorded.

they have a good degree of accuracy and are good measures of soil moisture, albeit in a surrogate form. Estimating the amount of water beneath the surface can also be carried out using either numerical modelling or remote sensing techniques. The main groundwater modelling techniques focus on the movement of water in the subsurface zone, using different forms of Darcy's law (e.g. Richards equation). There are also models of soil water balance that rely on calculating inflows (infiltration from rainfall) and outflows (seepage and evaporation) to derive a soil water storage value for a given time and space.

Remote sensing of soil moisture

In the last decade, significant advances in the field of remote sensing have allowed for increasingly accurate soil moisture measurements to be obtained by satellites. Both active and passive sensors can provide valuable information of soil moisture at medium-high resolution. The advantage of any remote sensing technique is that it samples over a wide spatial area. Satellites measure the electromagnetic radiation within each pixel; this is an average value for the whole area, rather than a point measurement that might be expected from normal soil moisture measurements. The question that needs to be answered before satellite remote sensing is widely accepted in hydrology is whether the enhanced spatial distribution of measurement is sufficient to overcome the limitations of direct soil moisture measurement. The key limitation being that to date, sensors can only produce accurate soil moisture measurements at the very near surface (i.e. within the top 5 cm), which is a restriction on their application in hydrological science. However, this is an important area in the generation of runoff (see Chapter 7) and is still worthy of measurement.

Passive sensors

Passive sensors rely on natural light to produce multispectral imagery (MSI) of the Earth's surface. While the common primary use of MSI is high resolution landcover mapping and the monitoring of vegetation/water conditions, it is possible to make estimates of soil moisture using MSI. For example, studies have shown that where soil moisture is within the range available to plants (i.e. plant available water) vegetation indices, such as the Moisture Stress Index (MSI) and Normalised Difference Vegetation Index (NDVI), are highly correlated with measured soil moisture in the surface horizons (Wang et al. 2007; Gu et al. 2008). As a result, NDVI from a range of sensors are key inputs into national drought monitoring systems, such as the US Drought Monitor (Brown et al. 2008) and the African Flood and Drought Monitor (Sheffield et al. 2014).

Beyond the multispectral region of the electromagnetic spectrum (EMS), thermal remote sensing also offers opportunities for the measurement of soil moisture. The high heat capacity of water means that it has considerable effect on the emission of thermal infrared signals from the earth. These can be detected by satellites and an inference made about how wet the soil is. This is especially so if two images of the same scene can be compared in order to derive a relative wetness. Satellite platforms like Landsat and SPOT are able to use this technique at spatial resolutions of around 10–30 m.

The major limitation with any passive remote sensing approaches is that they rely on a lack of cloud cover over the site of investigation, something that is not easy to guarantee, especially in hydrologically active (i.e. wet) areas.

Active sensors

Active sensors emit microwaves from a satellite and the strength of returning signal measured. This is a complex radar system and has only been available on satellites since the early 1990s. The strength of microwave backscatter is primarily dependent on two factors. Firstly soil roughness; where soil roughness is well known and there is little or no vegetation cover the radar backscatter has been well correlated with surface moisture (Griffiths and Wooding 1996; Kelly et al. 2003). Soil moisture content within the surface zone also influences the backscattering of radiation from a sensor as the soil dielectric properties change with variations in water content.

Commonly applied satellites which produce soil moisture measurements are the European Remote Sensing platforms (ERS-1 and ERS-2), the METeorological OPerational satellite (METOP), and the Soil Moisture Active Passive (SMAP) satellite. Launched in early 2015, SMAP was well positioned to revolutionise the field of soil moisture remote sensing through its unique combination of passive and active sensors with high spatial and temporal resolution. NASA hoped that SMAP would allow scientists to further our understanding of processes that link the terrestrial energy, water and carbon cycles. However only months into its 3-year mission, the radar on board SMAP suddenly stopped operating and was deemed unrepairable by ground teams. On launch there was an expectation that SMAP would produce both 9-km and 36-km global soil moisture datasets. With the loss of the radar, only the 36-km dataset is currently available. This dataset (and others from the SMAP mission) are available to download here: https://nsidc.org/data/smap/smap-data.html. There are plans to fuse SMAP datasets with those produced by other radar satellites (such as the European Space Agency Sentinel-1) to create a global scale, high resolution soil moisture product.

Case study

REMOTE SENSING OF SOIL MOISTURE AS A REPLACEMENT FOR FIELDWORK

Remote sensing of soil moisture may offer a way of deriving important hydrological information without intensive, and costly, fieldwork programmes. Grayson et al. (1992) suggest that this could be used to set the initial conditions for hydrological modelling, normally a huge logistical task. One of the major difficulties in this is that the accuracy of information derived from satellite remote sensing is not high enough for use in hydrological modelling. As a counter to this it can be argued that the spatial discretisation offered by remote sensing measurements is far better than that available through traditional field measurement techniques.

In an attempt to reconcile these differences, Davie et al. (2001) intensively monitored a 15-hectare field in eastern England and then analysed the satellite-derived, active microwave backscatter for the same period. The field programme consisted of measuring surface soil moisture (gravimetric method) at three different scales in an attempt to spatially characterise the soil moisture. The three scales consisted of: (a) 13 samples taken 1 m apart; (b) lines of samples 30 m apart; and (c) lines approximately 100 m apart. The satellite data was from the European Remote Sensing Satellite (ERS) using Synthetic Aperture Radar (SAR).

Analysis of the field data showed great variation in the surface soil moisture and a difference in measurement depending on the scale of measurement (see Figure 6.18). It is evident from Figure 6.18 that the point measurements of soil moisture are highly variable and that many measurements of soil moisture are required to try to characterise the overall field surface soil moisture.

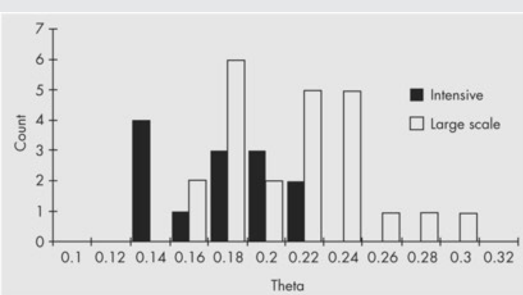


Figure 6.18 Measured surface soil moisture distributions at two different scales for a field in eastern England in October 1995.

In contrast to the directly measured soil moisture measurements, the microwave backscatter shows much less variation in response (see Figure 3, p. 330 of Davie et al. 2001). The most likely explanation for this is that the spatial averaging of backscatter response within a pixel (25 × 25 m in this case) evens out the variations found through point measurements. This can be thought of as a positive aspect because in many cases for hydrological modelling the scale needed is larger than point measurements and the spatial integration of backscatter response may be a way of getting around highly variable point measurements. When further analysis of the backscatter response was carried out it showed that interpretation at the pixel size was meaningless and they needed to be averaged-up themselves. It was found that the smallest resolution to yield meaningful results was around 1 hectare (100 × 100 m).

As with other studies on bare fields (e.g. Griffiths and Wooding 1996) there was a reasonably good correlation between radar backscatter and measured soil moisture. Unfortunately, this relationship is not good enough to provide more than around 70 per cent accuracy on estimations of soil moisture. In addition to this the study was carried out in conditions ideally suited for SAR interpretation (flat topography with no vegetation cover) which are far from atypical conditions encountered in hydrological investigations.

Overall it is possible to say that although the satellite remote sensing of soil moisture using SAR may offer some advantages of spatial integration of the data it is not enough to offset the inaccuracy of estimation, particularly in non-flat, vegetated catchments. There may come a time when satellite remote sensing can be used as a replacement for field measurements, but this will be in the future.

ESSAY QUESTIONS

1. **Compare and contrast different methods for measuring soil water content at the hillslope scale.**
2. **Define, compare and contrast terms *saturated and unsaturated hydraulic conductivity* and explain their importance in understanding movement of water in the ground.**
3. **Explain the fundamental forces influencing the movement of water in soil and why soil water is more difficult to model than groundwater.**
4. **Discuss the relevance of remote sensing for understanding soil moisture dynamics and consider the challenges of scale.**

FURTHER READING

Cooper, J.D. (2016) *Soil water measurement in the field: A practical handbook*. Wiley-Blackwell, Chichester.
An excellent new text focusing on measurement techniques.

Ferguson, R.I. (1999) Snowmelt runoff models. *Progress in Physical Geography* 23:205–228.
An overview of snow melt estimation techniques.

Freeze, R.A. and Cherry J.A. (1979) *Groundwater*. Prentice-Hall, Englewood Cliffs, N.J.
A classic textbook on groundwater (including soil water).

Hillel, D. (1998) *Environmental soil physics: Fundamentals, applications and environmental considerations*. Academic Press, San Diego.
A classic technical text.

Kendall, C. and McDonnell J.J. (eds) (1998) *Isotope tracers in catchment hydrology*. Elsevier Science, Amsterdam.
Detailed book on groundwater and streamwater ageing techniques.

Klute, A. (ed.) (1986) *Methods of soil analysis. Part 1: Physical and mineralogical methods*. American Society of Agronomy–Soil Science Society of America, Madison, W.I.
A mine of information on soil methods.

White, R. (2006) *Principles and practice of soil science: The soil as a natural resource* (4th edition) Blackwell Publishing, Malden, M.A.
An excellent text covering the discipline of soils science, including a chapter focussing on soil water.

7

RUNOFF

LEARNING OBJECTIVES

When you have finished reading this chapter you should have:

- An understanding of the process of runoff leading to **channel flow**.
- An understanding of how these processes determine the shape of the hydrograph and the catchment's response to rainfall.
- An awareness of the techniques for measuring hillslope runoff.
- Some insight into floods, flood causing mechanisms and impacts.

The amount of water within a river or stream is of great interest to hydrologists. It represents the end-product of all the other processes in the hydrological cycle and is where the largest amount of effort has gone into analysis of historical records. The methods of analysis are covered next in Chapter 9, whereas this chapter deals with the mechanisms that lead to water entering the stream: the runoff mechanisms. *Runoff* is a loose term that covers the movement of water to a channelised stream, after it has reached the ground as precipitation. The movement can occur either on or below the surface and at differing velocities. This is an important point; most people think of runoff in the sense of *surface* runoff, but the term includes water moving downslope within the soil profile, and of course groundwater flow (Chapter 5). Once the water reaches a stream it moves towards the oceans in a channelised form, the process referred to as **streamflow**, channel flow or **riverflow**. Streamflow is expressed as **discharge**: the volume of water over a defined time period. The SI units for discharge are m^3/s (colloquially referred to as *cumecs*). A continuous record of streamflow is called a **hydrograph** (see Figure 7.1). Although we think of this as continuous measurement, it is normally either an averaged flow over a time period or a series of samples (e.g. a measurement every 15 minutes). Often the y-axis of a hydrograph is plotted on a logarithmic scale. If hydrographs are plotted on a normal scale, showing the less frequent but high flow events means that much of the 'normal' lower flows are plotted very close to the x-axis

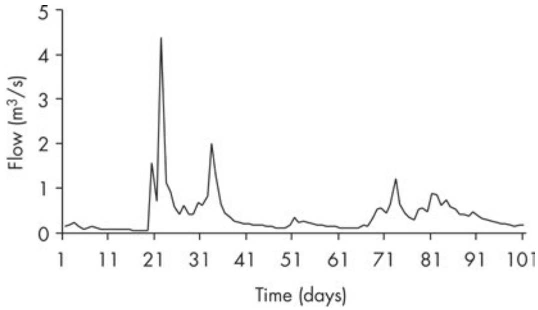


Figure 7.1 A typical hydrograph, taken from the river Wye, Wales for a 100-day period during the autumn of 1995. The values plotted against time are mean daily flow in cumecs.

and are difficult to see. Taking the logarithm of the flow and plotting that, allows the richness of the hydrograph to be visualised.

In Figure 7.1, there are a series of peaks between periods of steady, much lower flows. The hydrograph peaks are referred to as **peakflow**, **stormflow** or even **quickflow**. They are the water in the stream during and immediately after a significant rainfall event. The steady periods between peaks are referred to as **baseflow** or sometimes delayed flow or slowflow (N.B. this is different from **low flow**; see Chapter 9).

The shape of a hydrograph, and in particular the shape of the stormflow peak, is influenced by the storm characteristics (e.g. rainfall intensity and duration) and many physical characteristics of the upstream catchment. In terms of catchment characteristics, the largest influence is exerted by catchment size, but other factors include slope, shape of catchment, soil type, geology, vegetation type and percentage cover, **hydraulic length**, **drainage density**, degree of urbanisation and the antecedent soil moisture. The significant effect catchment characteristics can have on streamflow is illustrated for the Lambourne at Shaw and the Ock at Abingdon in Figure 7.2. Both these are similarly sized ($234km^2$) sub-catchments of the Thames, UK – and as directly neighbouring catchments, they share a virtually identical climate (see data for each station at the National River Flow Archive (CEH 2015), https://nrfa.ceh.ac.uk/). What differs principally is their underlying geology. The rocks of the Lambourn are virtually all high permeability chalk, whilst the Ock has some chalk but also bedrock of lower permeabilities.

Figure 7.3 shows the shape of a storm hydrograph in detail. There are several important hydrological terms that can be seen in this diagram. The **rising limb** of the hydrograph is the initial steep part leading up to the highest or peakflow value. The water contributing to this part of the hydrograph is from rapid runoff mechanisms and *channel precipitation* (i.e. rain that falls directly onto the channel). Some texts claim that channel precipitation shows up as a preliminary blip before the main rising limb. In reality this is very rarely observed, a factor of the complicated nature of storm runoff processes. The **recession limb** of the hydrograph is after the peak and is characterised by a longer, slower decrease in streamflow until the baseflow is reached again. The recession limb is attenuated by two factors: storm water arriving at the mouth of a catchment from the furthest parts, and the arrival of water that has moved as subsurface flow at a slower rate than the streamflow. The time difference between the peakflow value and the rainfall event driving the flow is the lag and is also a function of the same catchment physical properties that determine the shape of the hydrograph. The area under the curve above the dotted baseflow line represents the stormflow volume (m^3).

Exactly how water moves from precipitation reaching the ground surface to channelised streamflow is one of the most intriguing hydrological questions, and one that cannot be answered easily. Much research effort in the past hundred years has gone into understanding runoff mechanisms; considerable advances have been made, but there are still many unanswered questions. The following section describes how it is thought runoff occurs, but there are many different scales at which these mechanisms are evident and they do not occur everywhere.

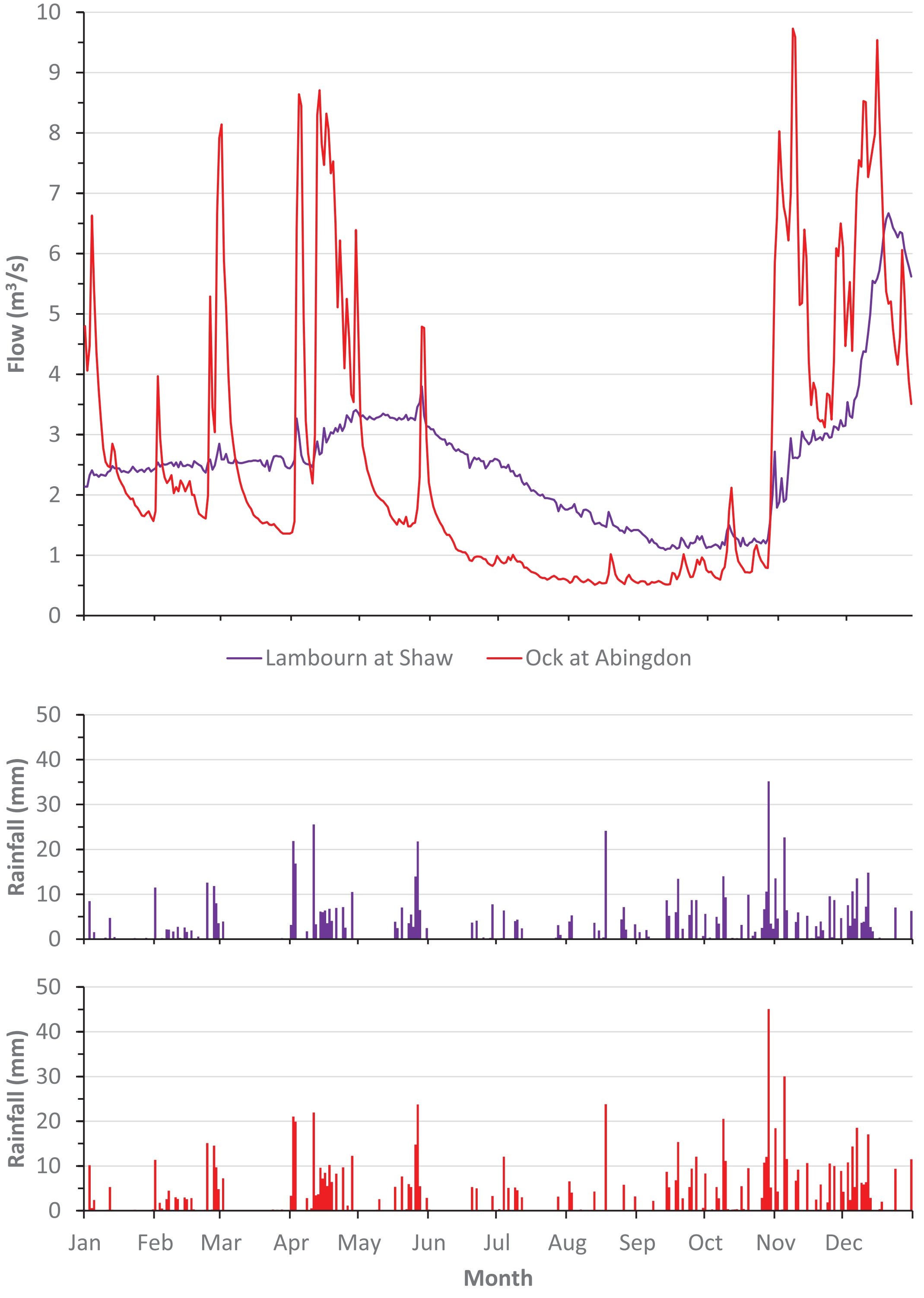


Figure 7.2 Comparative hydrographs for two adjacent sub-catchments in the Thames catchment with near identical climatic conditions but with different geology. The values plotted against time are mean daily flow (m^3/s).

Source: Data from the National River Flow Archive, Centre for Ecology and Hydrology, United Kingdom

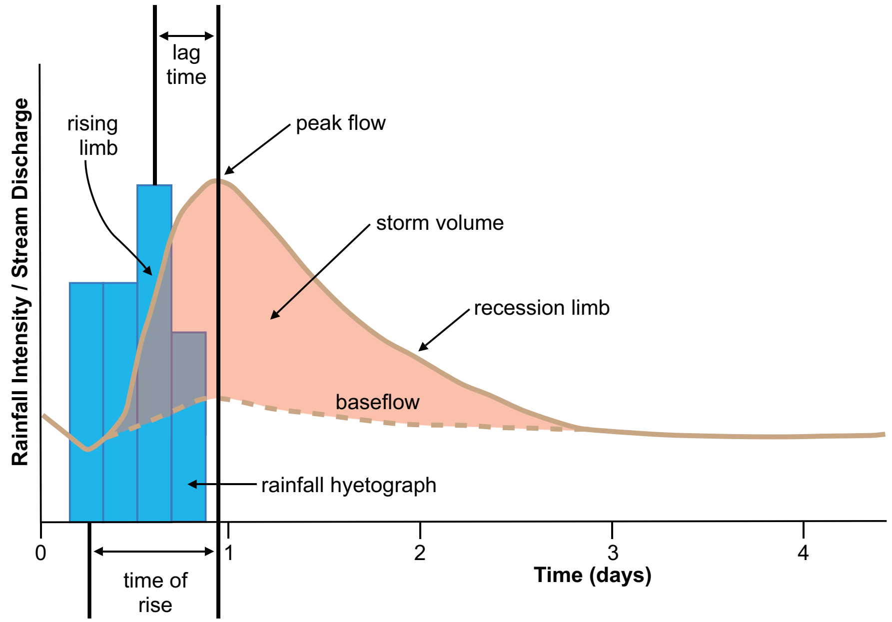


Figure 7.3 A schematic storm hydrograph.

RUNOFF MECHANISMS

Figure 7.4 is an attempt to represent the different runoff processes that can be observed at the hillslope scale. **Overland flow** (Q_o) is the water which runs across the surface of the land before reaching the stream. In the subsurface, throughflow (Q_t) (some authors refer to this as **lateral flow** or **transverse flow**) occurs in the shallow subsurface, predominantly, although not always, in the unsaturated zone. Groundwater flow (Q_G) is in the deeper saturated zone. All of these are runoff mechanisms that contribute to streamflow. The relative importance of each is dependent on the catchment under study and the rainfall characteristics during a storm.

Overland flow

Some of the earliest research work on how overland flow occurs was undertaken by Robert Horton (1875–1945). In a classic paper from 1933, Horton hypothesised that overland flow occurred when the rainfall rate was higher than the infiltration rate of a soil. Horton went on to suggest that under these circumstances the excess rainfall collected on the surface before travelling towards the stream as a thin sheet of water moving across the surface. Under this hypothesis it is the infiltration rate of a soil that acts as a controlling barrier or partitioning device. Where the infiltration capacity of a soil is low, overland flow occurs readily. This type of overland flow is referred to as **infiltration excess overland flow** or **Hortonian overland flow** although as Beven (2004) points out, Horton himself referred to it as 'rainfall excess'. Horton's view that the storm response of catchments was dominated by overland flow remained the dominant theory for decades (Bevan 2006).

Horton's ideas were extremely important in hydrology as they represented the first serious attempt to understand the processes of storm runoff that lead to a storm hydrograph. However, when people started to measure infiltration capacities of soils they invariably found that they were far higher

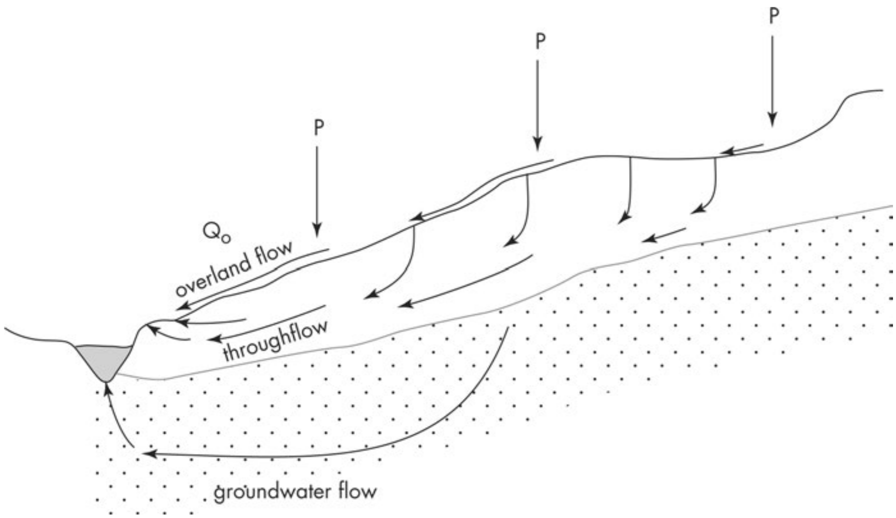


Figure 7.4 Hillslope runoff processes. See text for explanation of terms.

Source: Adapted from Dunne (1978)

than most normal rainfall rates. This is illustrated in Table 7.1 where some typical infiltration capacities and rainfall rates are shown. Other measurements confirm high infiltration capacities for soils, e.g. Selby (1970) reports infiltration capacities of between 60 and 600 mm/hour on short grazed pasture on yellow-brown pumice soils in the central North Island of New Zealand. The values were higher for ungrazed grass and under trees and are generally higher than the measured rainfall intensities (Selby 1970).

Table 7.1 Some typical infiltration rates compared to rainfall intensities

Soil and vegetation	*Infiltration rate (mm/hr)*	*Rainfall type*	*Rainfall intensity (mm/hr)*
Forested loam	100–200	Thunderstorm	50–100
Loam pasture	10–70	Heavy rain	5–20
Sand	3–15	Moderate rain	0.5–5
Bare clay	0–4	Light rain	0.5

Source: From Burt (1987)

In addition to the infiltration capacity information, it is extremely rare to find a thin sheet of water moving over the surface during a storm event. It was observations such as those by Hursh (1944) and others that subsequently led to a general revision of Horton's hypothesis. Hursh (1944), for example, recognised that flows other than surface flow might be important, and coined the term 'subsurface-stormflow'. His 1944 paper considered how subsurface flow might be relevant in the storm hydrograph (Bevan 2006). Cook (1946) proposed that overland flow could also be generated by subsurface flow that re-emerges onto the land surface, an idea later demonstrated in field experiments (Kirkham 1947). However, one of the first to question the dominance of Horton's hypothesis (at least in print – as Bevan (2006) points out) was Roger Betson. In his study catchments, Betson (1964) showed that storm runoff was being generated from relatively small parts of these catchments. His proposition was that within a catchment there are only limited (partial) areas that contribute overland flow to a storm hydrograph. This is referred to as the **partial areas concept**. Betson did not challenge the role of infiltration excess overland flow as the primary source of stormflow, but did challenge the idea of overland flow occurring as a thin sheet of water throughout a catchment.

Hewlett and Hibbert (1967) were the first to suggest that there might be another generally applicable mechanism of overland flow occurring. In contrast to Horton who worked in a more arid part of the USA, Hewlett and Hibbert were part of a group of forest hydrologists that had for some time been less convinced of the applicability of his theory in more temperate areas. This was particularly based on the observations from the eastern USA: that during a storm it was common to find all the rainfall infiltrating a soil. Hewlett and Hibbert (1967) hypothesised that during a rainfall event all the water infiltrated the surface. This hypothesis was confirmed by a comprehensive field study by Dunne and Black (1970).

Through a mixture of infiltration and throughflow, the water table would rise until in some places it reached the surface. At this stage overland flow occurs as a mixture of return flow (i.e. water that has been beneath the ground but returns to the surface) and rainfall falling on saturated areas. This type of overland flow is referred to as **saturated overland flow**. Hewlett and Hibbert (1967) suggested that the water table was closest to the surface, and therefore likely to rise to the surface quickest, adjacent to stream channels and at the base of slopes. Their ideas on stormflow were that the areas contributing water to the hydrograph peaks were the saturated zones, and that these vary from storm to storm. In effect the saturated areas immediately adjacent to the stream act as extended channel networks. This is referred to as the **variable source areas concept**. This goes a step beyond the ideas of Betson (1964) as the catchment has a partial areas response but the response area is dynamic; i.e. variable in space and time.

So who was right: Horton, or Hewlett and Hibbert? The answer is that both were. Table 7.2 provides a summary of the ideas for storm runoff generation described here. It is now accepted that saturated overland flow (Hewlett and Hibbert) is the dominant overland flow mechanism in humid, mid-latitude areas. It is also accepted that the variable source areas concept is the most valid description of stormflow processes. However, where the infiltration capacity of a soil is low or the rainfall rates are high, Hortonian overland flow does occur. In Table 7.1 it can be seen that there are times when rainfall intensities will exceed infiltration rates under natural circumstances. It is also important to note that these processes could be occurring at the same time in the same catchment. In arid and semi-arid zones it is not uncommon to find extremely high rainfall rates (fed by convective storms) that can lead to infiltration excess overland flow and rapid flood events; this is called **flash flooding**.

Examples of low infiltration rates can be found with compacted soils (e.g. from vehicle movements in an agricultural field), on roads and paved areas, on heavily crusted soils and what are referred to as **hydrophobic soils**.

Basher and Ross (2001) reported infiltration capacities of 400 mm/hour in market gardens in the North Island of New Zealand and that these rates increased during the growing season to as high as 900 mm/hour. However, Basher and Ross (2001) also showed a decline in infiltration

Table 7.2 A summary of the ideas on how stormflow is generated in a catchment

	Horton	*Betson*	*Hewlett and Hibbert*
Infiltration	Controls overland flow	Controls overland flow	All rainfall infiltrates
Overland flow mechanism	Infiltration excess	Infiltration excess	Saturated overland flow
Contributing area	Uniform throughout the catchment	Restricted to certain areas of the catchment	Contributing area is variable in time and space

capacity to as low as 0.5 mm/hour in wheel tracks at the same site.

Hydrophobic soils have a peculiar ability to swell rapidly on contact with water, which can create an impermeable barrier at the soil surface to infiltrating water, leading to Hortonian overland flow. The cause of hydrophobicity in soils has been linked to several factors including the presence of mycorrhizal fungi and swelling clays such as allophane (Doerr et al. 2007). Hydrophobicity is a temporary soil property; continued contact with water will increase the infiltration rate. For example Clothier et al. (2000) showed how a yellow brown earth/loam changed from an initial infiltration capacity of 2 mm/hour to 14 mm/hour as the soil hydrophobicity breaks down.

In Hewlett and Hibbert's (1967) original hypothesis it was suggested that contributing saturated areas would be immediately adjacent to stream channels. Subsequent work by the likes of Dunne and Black (1970), Anderson and Burt (1978) and others has identified other areas in a catchment prone to inducing saturated overland flow. These include areas where there is convergence of flow e.g. hillslope hollows, slope concavities (in section) (Figure 7.5a), or where there is a thinning of the soil overlying an impermeable base (Figure 7.5c). Similarly where a change of slope results in a reduction of hydraulic gradient (Figure 7.5b) or a change in hydraulic conductivity (Figure 7.5d), any throughflow is likely to return to the surface as the volume of soil receiving it is not large enough for the amount of water entering it. This can be commonly observed in the field where wet and boggy areas can be found at the base of slopes and at valley heads (hillslope hollows). Because these areas may not be

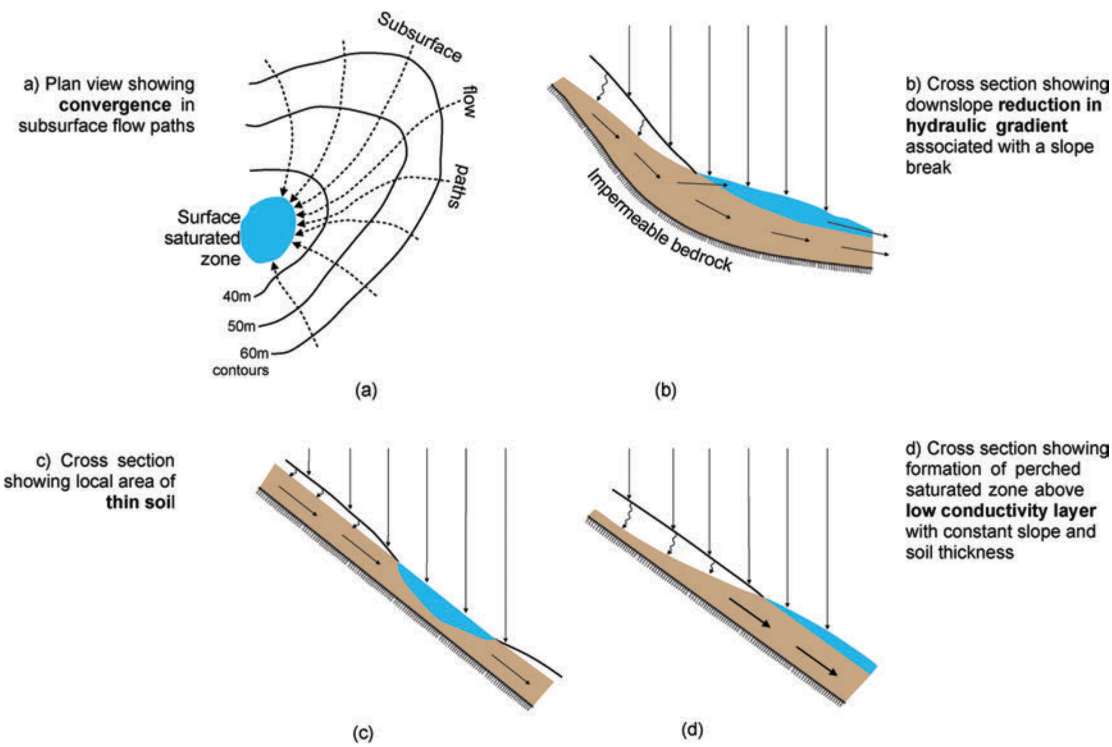


Figure 7.5 Potential disjunct source areas.

Source: Adapted from Dingman (2008) and Ward and Robinson (2000) immediately adjacent to streamside variable source areas they are called **disjunct source areas**, but for them to contribute to stormflow, they would need to have effective hydrological connections with valley bottoms (Robinson and Ward 2017).

Subsurface flow

Under the variable source areas concept there are places within a catchment that contribute overland flow to the storm hydrograph. When we total up the amount of water found in a storm hydrograph it is difficult to believe that it has all come from overland flow, especially when this is confined to a relatively small part of the catchment (i.e. variable source areas concept). The more gradual manner in which the recession limb of a hydrograph attenuates the storm-flow suggests that it may be derived from a slower movement of water: subsurface flow. In addition to this, tracer studies looking at where the water has been before entering the stream as stormflow have found that a large amount of the storm hydrograph consists of 'old water' (e.g. Martinec et al. 1974; Fritz et al. 1976). This old water has been sitting in the soil, or as fully saturated groundwater, for a considerable length of time and yet enters the stream during a storm event. There have been several theories put forward to try and explain these findings, almost all involving throughflow and groundwater.

Throughflow is a general term used to describe the movement of water through the unsaturated zone; normally this is the soil matrix. Once water infiltrates the soil surface it continues to move, either through the soil matrix or along preferential flow paths (referred to as lateral or preferential flow). The rate of soil water movement through a saturated soil matrix is described by Darcy's law (see Chapter 5) and the Richards approximation of Darcy's law when below saturation. Under normal, vertical, infiltration conditions the hydraulic gradient has a value of –1 and the saturated hydraulic conductivity is the infiltration capacity. Once the soil is saturated, the movement of water is not only vertical. With a sloping water table on a hillslope, water moves down slope. However, the movement of water through a saturated soil matrix is not rapid, e.g. Kelliher and Scotter (1992) report a K_{sat} value of 13 mm/hour for a fine sandy loam. In order for throughflow to contribute to storm runoff there must be another mechanism (other than matrix flow) operating.

One of the first theories put forward concerning the contribution of throughflow to a storm hydrograph was by Horton and Hawkins (1965) (this Horton was a different person from the proposer of Hortonian overland flow). They proposed the mechanism of *translatory* or *piston flow* to explain the rapid movement of water from the subsurface to the stream. They suggested that as water enters the top of a soil column it displaces the water at the bottom of the column (i.e. old water), and the displaced water enters the stream. The analogy is drawn to a piston where pressure at the top of the piston chamber leads to a release of pressure at the bottom. The release of water to the stream can be modelled as a pressure wave rather than tracking individual particles of water. Piston flow has been observed in laboratory experiments with soil columns (e.g. Germann and Beven 1981).

At first glance the simple piston analogy seems unlike a real-life situation, since a hillslope is not bounded by impermeable sides in the same way as a piston chamber. However, the theory is not as farfetched as it may seem, as the addition of rainfall infiltrating across a complete hillslope is analogous to pressure being applied from above and in this case the boundaries are upslope (i.e. gravity) and the bedrock below. Brammer and McDonnell (1996) suggest that this may be a mechanism for the rapid movement of water along the bedrock and soil interface on the steep catchment of Maimai in New Zealand. In this case it is the hydraulic gradient created by an addition of water to the bottom of the soil column, already close to saturated, that forces water along the base where hydraulic conductivities are higher.

Ward (1984) draws the analogy of a thatched roof to describe the contribution of subsurface flow to a stream (based on the ideas of Zaslavsky and Sinai (1981)). When straw is placed on a sloping roof it is very efficient at moving water to the bottom of the roof (the guttering being analogous to a stream) without visible overland flow. This is due to the preferential flow direction along, rather than between, sloping straws. Measurements of hillslope soil properties do show a higher hydraulic conductivity in the downslope rather than vertical direction. This would account for a movement of water downslope as throughflow, but it is still bound up in the soil matrix and reasonably slow.

There is considerable debate on the role of **macropores** in the rapid movement of water through the soil matrix. Macropores are larger pores within a soil matrix, typically with a diameter greater than 3 mm. They may be caused by soils cracking, worms burrowing or other biotic activities. The main interest in them from a hydrologic point of view is that they provide a rapid conduit for the movement of water through a soil. The main area of contention concerning macropores is whether they form continuous networks allowing rapid movement of water down a slope or not. There have been studies suggesting macropores as a major mechanism contributing water to stormflow (e.g. Mosley 1979, 1982; Wilson et al. 1990), but it is difficult to detect whether these are from small areas on a hillslope or continuous throughout. Jones (1981) and Tanaka (1992) summarise the role of pipe networks (a form of continuous macropores) in hillslope hydrology. Where found, pipe networks have considerable effect on the subsurface hydrology but they are not a common occurrence in the field situation.

The role of macropores in runoff generation is unclear. Although they are capable of allowing rapid movement of water towards a stream channel there is little evidence of networks of macropores moving large quantities of water in a continuous fashion. Where macropores are known to have a significant role is in the rapid movement of water to the saturated layer (e.g. Heppell et al. 1999) which may in turn lead to piston flow (McGlynn et al. 2002).

Groundwater contribution to stormflow

Another possible explanation for the presence of old water in a storm hydrograph is that it comes from the saturated zone (groundwater) rather than from throughflow. This is contrary to conventional hydrological wisdom which suggests that groundwater contributes to baseflow but not to the stormflow component of a hydrograph. Although a groundwater contribution to stormflow had been suggested before, it was not until Sklash and Farvolden (1979) provided a theoretical mechanism for this to occur that the idea was seriously considered. They proposed the capillary fringe hypothesis to explain the groundwater ridge, a rise in the water table immediately adjacent to a stream (as observed by Ragan 1968). Sklash and Farvolden (1979) suggested that the addition of a small amount of infiltrating rainfall to the zone immediately adjacent to a stream causes the soil water to move from an unsaturated state (i.e. under tension) to a saturated state (i.e. a positive pore pressure expelling water). As explained in Chapter 6, the relationship between soil water content and soil water tension is non-linear. The addition of a small amount of water can cause a rapid change in soil moisture status from unsaturated to saturated. This provides the groundwater ridge which:

> not only provides the early increased impetus for the displacement of the groundwater already in a discharge position, but it also results in an increase in the size of the groundwater discharge area which is essential in producing large groundwater contributions to the stream.
>
> (Sklash and Farvolden 1979: 65)

An important point to stress from the capillary fringe hypothesis is that the groundwater ridge is developing well before any throughflow may have been received from the contributing hillslope areas. These ideas confirm the variable source areas concept and provide a mechanism for a significant old water contribution to storm hydrographs. Field studies such as that by McDonnell (1990) have observed groundwater ridging to a limited extent, although it is not an easy task as often the instrument response time is too slow to detect the rapid change in pore pressure properly.

Case study

THE MAIMAI RUNOFF GENERATION STUDIES

The Maimai catchment study (near Reefton on the west coast of the South Island of New Zealand) was established in 1974 for research into the effects of logging native beech forest (*Nothofagus*) and replanting with different non-indigenous species (Figure 7.6). The installation of hydrological measuring equipment and the fact that rainfall and stormflow are frequently observed made it an ideal place for studying stormflow generation mechanisms in depth. The knowledge gained from detailed hydrological process studies at Maimai have played a major part in shaping thinking on stormflow generation mechanisms.

The Maimai catchment is characterised by short, steep slopes (approximately 300 m with angles of around 35°), covered in thick vegetation, with incised channels and very small valley bottoms. Annual rainfall is approximately 2,600 mm with an average of 156 rain days a year, and stormflow makes up 65 per cent of the total streamflow (Rowe et al. 1994; Pearce et al. 1986).

Figure 7.6 Maimai catchments in South Island, New Zealand. At the time of the photograph (1970s) five catchments had been logged and were about to be replanted with *Pinus radiata*.

Mosley (1979, 1982) used Maimai to investigate the role of macropores as conduits for rapid movement of rainfall to the stream. Observations of macropore flow rates using cut soil faces and dye tracers suggested that rainfall could travel down the short steep hillslopes at Maimai in less than 3 hours (i.e. within the time frame of a storm event). Subsequent chemical and isotopic analysis of streamflow, rainfall and water exiting the cut soil pit faces showed that the majority of measured streamflow was 'old' water, suggesting that rapid, extensive macropore flow was not the main mechanism for stormflow generation (Pearce et al. 1986).

McDonnell (1990) investigated this further, in particular looking at possible groundwater ridging (Sklash and Farvolden 1979) as a mechanism for large amounts of old water as saturated overland flow. Although this could be observed at Maimai, the amount of water held near the stream prior to an event was not large enough to account for all of the old water, which suggested that another mechanism (e.g. piston flow) might be working (McDonnell 1990).

McGlynn et al. (2002) present a summary conceptual diagram of runoff mechanisms on Maimai hillslopes that combines many of the features described above (see Figure 7.7). In this model there is rapid infiltration of water through macropores to reach the bedrock. At this stage a form of piston flow occurs as the saturated zone at the base of the soil mantle is confined by the soil matrix above it. At the bedrock interface there may be a network of macropores or else the same situation of a confined aquifer in that the soil matrix above has a much lower hydraulic conductivity. Water is then pushed out at the bottom due to the pressure from new water arriving directly at the bedrock interface. There is also a mixing of the new water with old water sitting in bedrock hollows, creating a rapid movement of old water into the stream during storm events.

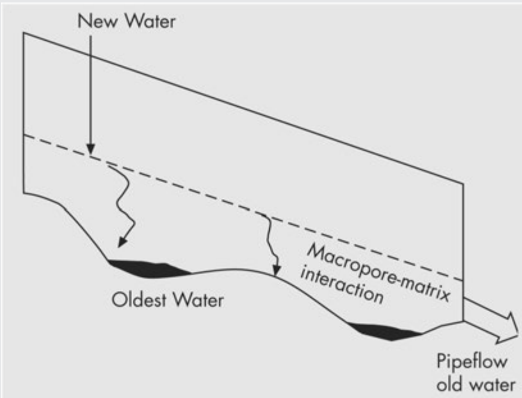


Figure 7.7 Summary hypothesis for hillslope stormflow mechanisms at Maimai. Rapid movement of water occurs through rapid infiltration to the bedrock interface and then a form of piston flow along this interface.

Source: Adapted from McGlynn et al. (2002)

How relevant are the Maimai stormflow generation studies?

The studies that have taken place at Maimai have been extremely important in influencing hydrological thinking around the world. However, an argument can be made that the conditions at Maimai are far from generally applicable elsewhere. The main study catchment (M8) has short, steep slopes and is in an area of high, and frequent, rainfall. The soils are extremely porous (infiltration rates in excess of 1,600 mm/hour have been measured) and remain within 10 per cent of saturation for most of the year (Mosley 1979). These conditions are not common and it would be difficult to generalise the concepts beyond Maimai. One of the really important concepts that Maimai has shown is that under conditions ideal for stormflow generation the mechanisms are still extremely complex and spatially variable. This is true wherever in the world the study is taking place.

Summary of storm runoff mechanisms

The mechanisms that lead to a storm hydrograph are extremely complex and still not fully understood. Although this would appear to be a major failing in a science that is concerned with the movement of water over and beneath the surface, it is also an acknowledgement of the extreme diversity found in nature. In general there is a reasonable understanding of possible storm runoff mechanisms but it is not possible to apply this universally. In some field situations the role of throughflow and piston flow are important, in others not; likewise for groundwater contributions, overland flow and pipeflow. Figure 7.8 provides a conceptual overview of the types of flow processes in relation to the potential contribution to the overall hydrograph. Note that the boundaries are purposefully blurred. Clearly, rapid response mechanisms are responsible for the peak, and the slower are responsible for the recession, but specific local characteristics will determine the 'hydrological signature' of the particular catchment. The challenge for modern hydrology is to identify quickly the specific dominant mechanisms for a particular hillslope or catchment so that the understanding of the hydrological processes in that situation can be used to aid management of the catchment.

The processes of storm runoff generation described here are mostly observable at the hillslope scale, and different processes will be occurring simultaneously throughout a catchment (Figure 7.9). At the catchment scale (and particularly for large river basins) the timing of peak flow (and consequently the shape of the storm hydrograph) is influenced more by the channel drainage network and the precipitation characteristics of a storm than by the mechanisms of runoff. This is a good example of the problem of scale described in Chapter 1. At the small hillslope scale storm runoff generation mechanisms are important, but they become considerably less so at the much larger catchment scale.

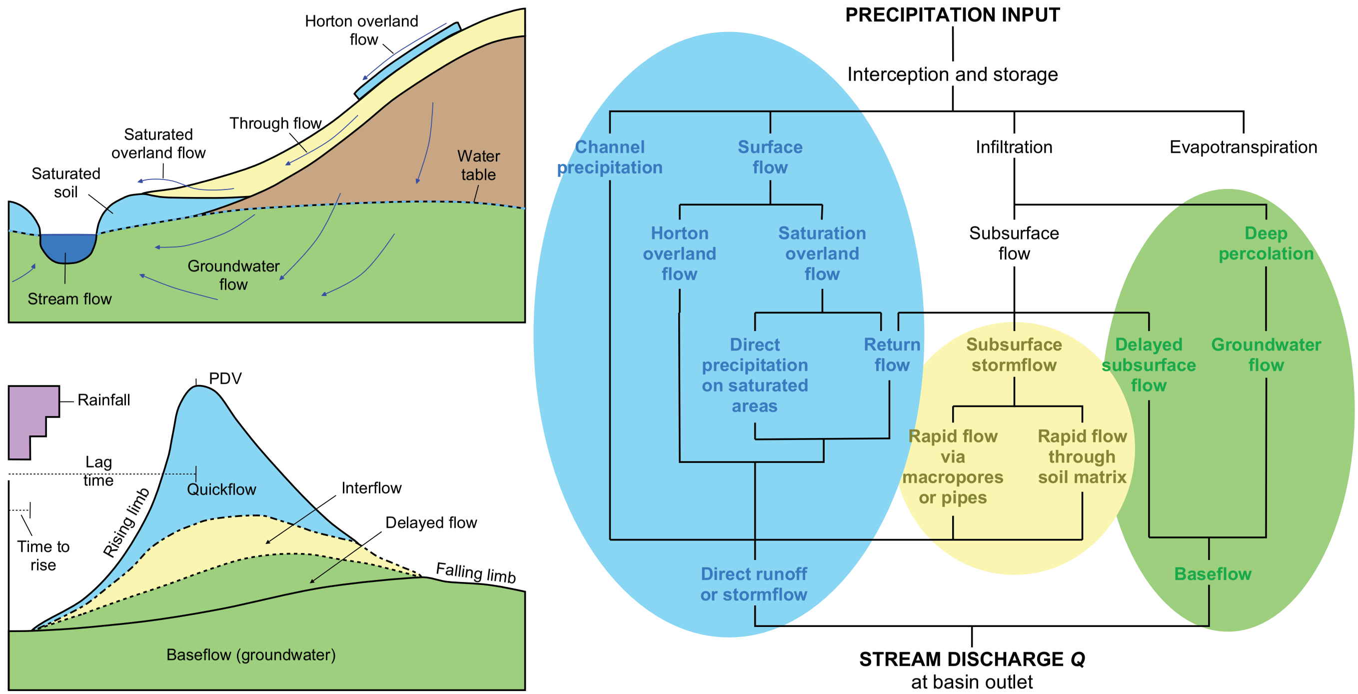


Figure 7.8 Runoff generation processes in relation to the generated hydrograph.

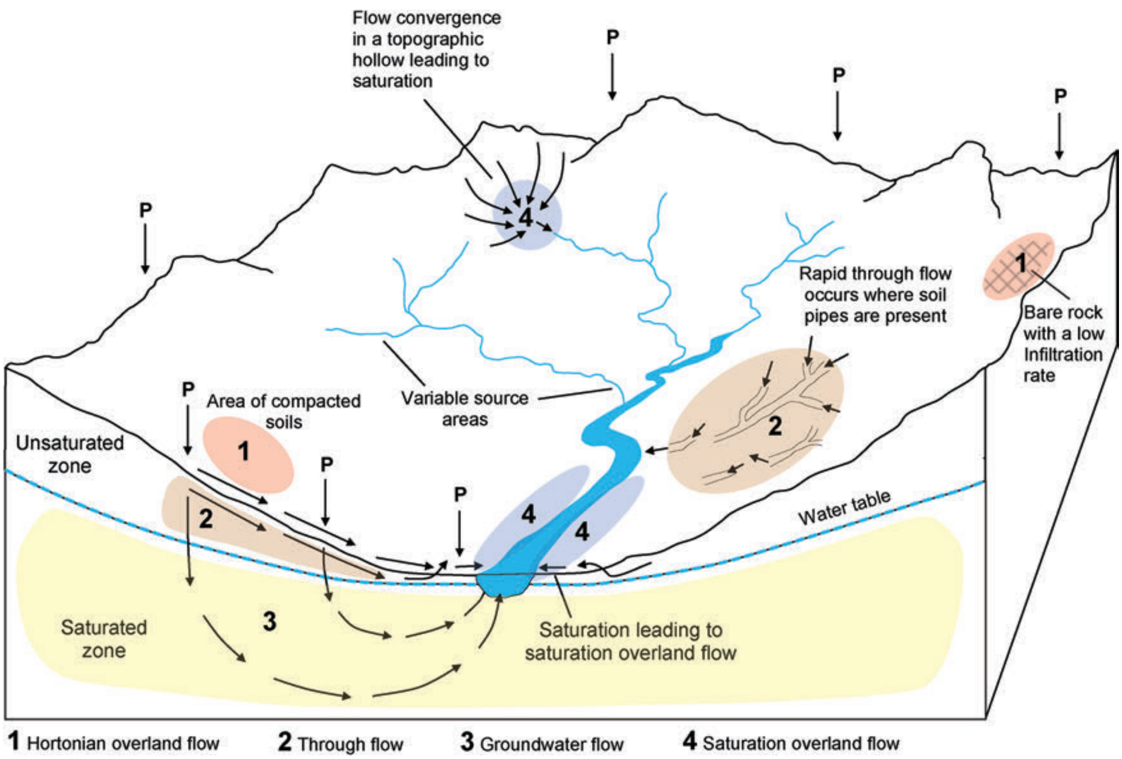


Figure 7.9 Runoff generation processes occurring throughout a catchment.

Source: Reproduced from Charlton (2008), with permission

Baseflow

In sharp contrast to the storm runoff debate, there is general consensus that the major source of baseflow is groundwater – and to a lesser extent throughflow. This is water that has infiltrated the soil surface and moved towards the saturated zone. Once in the saturated zone it moves downslope, often towards a stream. A stream or lake is often thought to occur where the regional water table intersects the surface, although this may not always be the case. In Chapter 5 the relationship between groundwater and streamflow has been explained (see Figure 5.5). However, in general it can be said that baseflow is provided by the slow seepage of water from groundwater into streams. This will not necessarily be visible (e.g. springs) but can occur over a length of streambank and bed and is only detectable through repeated measurement of streamflow down a reach.

Channel flow

Once water reaches the stream it will flow through a channel network to the main river. The controls over the rate of flow of water in a channel are to do with the volume of water present, the gradient of the channel, and the resistance to flow experienced at the channel bed. This relationship is described in uniform flow formulae such as the Chezy and Manning equations (see p. 173). The resistance to flow is governed by the character of the bed surface. Boulders and vegetation will create a large amount of friction, slowing the water down as it passes over the bed.

In many areas of the world, the channel network is highly variable in time and space. Small channels may be ephemeral and in arid regions will frequently only flow during flood events. The resistance to flow under these circumstances is complicated by the infiltration that will be occurring at the water front and bed surface. The first flush of water will infiltrate at a much higher rate as it fills the available pore space in the soil/rock at the bed surface. This will remove water from the stream and also slow the water front down as it creates a greater friction surface. Under a continual flow regime the infiltration from the stream to ground will depend on the hydraulic gradient and the infiltration capacity.

Measuring hillslope runoff

The measurement of runoff may be required to assess the relative contribution of different hillslope runoff processes, i.e. throughflow, overland flow, etc. There are no standard methods for the measurement of runoff processes; different researchers use different techniques according to the field conditions expected and personal preference.

Overland flow

The amount of water flowing over the soil surface can be measured using collection troughs at the bottom of hillslopes or runoff plots. A runoff plot is an area of hillslope with definite upslope and side boundaries so that you can be sure all the overland flow is generated from within each plot. The upslope and side boundaries can be constructed by driving metal plates into the soil and leaving them protruding above the surface. All flow is therefore contained and can be directed to either a tipping bucket type instrument or temporary flume where it can be measured as a volume per unit time (see next chapter). It is normal to use several runoff plots to characterise overland flow on a slope as it varies considerably in time and space. This spatial and temporal variation may be overcome with the use of a rainfall simulator.

Throughflow

Measurement of throughflow is fraught with difficulty. The only way to measure it is with throughflow troughs dug into the soil at the depths of interest in the soil profile. For example, if one was measuring throughflow down a slope, a trench would need to be dug perpendicular to the flow direction. Flow at various depths would need to be contained and separated for measurement. The problem with this is that in digging, the soil profile is disturbed and consequently the flow characteristics change. It is usual to insert troughs into a soil face that has been excavated and then refill the hole. This may still overestimate throughflow as the reconstituted soil in front of the troughs may encourage flow towards it as an area allowing rapid flow. Subsurface flow tracing has also been used – this involves adding a conservative tracer (i.e. something that does not bind to the soil) to water to enable the flow path to be detected. Tracers include sodium chloride and bromide but also fluorescent dyes such as RhodamineWT (Shaw et al. 2011). These methods represent a very complex experimental setup (see Shaw et al. (2011) for a more complete description).

FLOODS

The term *flood* is difficult to define except in the most general of terms. In a river a flood is normally considered to be an inundation of land adjacent to a river caused by a period of abnormally large discharge (Figure 7.10 and 7.11) or encroachment by the sea, but even this definition is fraught with inaccuracy. Flooding may occur from sources other than rivers (e.g. the sea and lakes), and 'abnormal' is difficult to pin down, particularly within a timeframe. Floods come to our attention through the amount of damage that they cause and for this reason they are often rated on a cost basis rather than on hydrological criteria. Hydrological and monetary assessments of flooding often differ markedly because the economic valuation is highly dependent on location. If the area of land

Figure 7.10 A river in flood. The excess water has spread across the floodplain outside the main river channel.

Figure 7.11 Images of flood inundation in Fiji, 2007.

inundated by a flooding river is in an expensive region with large infrastructure then the cost will be considerably higher than, say, for agricultural land. Two examples of large-scale floods during the 1990s illustrate this point. In 1998 floods in China caused an estimated US$20 billion of damage with over 15 million people being displaced and 3,000 lives lost (Smith 2001). This flood was on a similar scale to one that occurred in the same region during 1954. A much larger flood (in a hydrological sense) in the Mississippi and Missouri rivers during 1993 resulted in a similar economic valuation of loss (US$15–20 billion) but only 48 lives were lost (USCE 1996). The flood was the highest in the hydrological record and had an average recurrence interval of between 100 and 500 years (USCE 1996). The difference in lives lost and relative economic loss (for size of flood) is a reflection of the differing response to the flood in two economically contrasting countries.

As described in Chapter 2 for precipitation, flooding is another example where the *frequency–magnitude relationship* is important. Small flood events happen relatively frequently whereas the really large floods occur rarely but cause the most damage. The methods for interpreting riverflows that may be used for flood assessment are discussed in Chapter 6. They provide some form of objective flood size assessment, but their value is highly dependent on the amount of data available.

Floods are a frequently occurring event around the world, for example in June and July 2007 there were eleven large flood events reported in the news media (see Table 7.3). These floods were caused by varying amounts of rainfall, and occurred in different seasons of the year but all caused significant damage and in many cases loss of lives. There are numerous reasons why a river will flood and they almost always relate back to the processes found within the hydrological cycle. The main cause of river floods is when there is too much rainfall for the river to cope with. Other, more special causes of floods are individual events like dam bursts, **jökulhlaups** (ice-dam bursts) or snow melt (see pp. 78–81).

Table 7.3 Flooding events in news reports during June–July 2007

Location (date)	*Rainfall or flood statistics*	*Effect*
Midlands and Yorkshire, UK (June 2007)	One location 103 mm of rainfall in 24 hours; many places recorded over 50 mm of rain in 12 hours	30,000+ houses affected; estimated £1.5 bn damage
New South Wales, Australia (June 2007)	300 mm rainfall in 3 days	Nine lives lost, 5,000 evacuated
Bangladesh (June 2007)	400 mm cumulative rainfall in places	130 lives lost, 10,000 evacuated
India (June 2007)	475 mm rainfall in 4 days	57 lives lost, 100,000 people evacuated
China (June–July 2007)	300 mm rainfall in 4 days	88 lives lost, 500,000 people evacuated; 56,000 homes destroyed; 91,800 ha crops destroyed
Mid-West, USA (July 2007)	305 mm rainfall in 7 days	17 lives lost
Pakistan (July 2007)	105 mm rainfall in 12 hours; 30-year record	110 lives lost, 200,000 homeless
Southern Japan (July 2007)	200 mm rainfall in 4 days	3,400 evacuated
Sudan (July 2007)	At several sites the Nile was more than 1 m higher than in 1988 (a previous record level)	59 lives lost; 30,000 homes evacuated
Northland, New Zealand (July 2007)	270 mm rain total; 213 mm rainfall in 24 hours; 1 in 150-year storm	23 houses destroyed. Estimated damage $80 m (≈ US$60 m)
Midlands, England, UK (July 2007)	121 mm of rainfall in 24 hours; wettest May–July since records began in 1766	Seven people killed, estimated £2 bn damage

Influences on flood size

The extent and size of the flood can often be related to other contributing factors that increase the effect of high rainfall. Some of these factors are described here but all relate back to concepts introduced in earlier chapters detailing the processes found within the hydrological cycle. Flooding provides an excellent example of the importance of scale, introduced in Chapter 1. Many of the factors discussed here have an influence at the small scale (e.g. hillslopes or small research catchments of less than 10 km^2) but not at the larger overall river catchment scale. This section explores some of the factors compounding flood magnitude and we consider two case studies; flooding in Mozambique in 2000, and in Brisbane in 2010/11.

Antecedent soil moisture

The largest influence on the size of a flood, apart from the amount and intensity of rainfall, is the wetness of the soil immediately prior to the rainfall or snow melt occurring. As described on p. 123, the amount of infiltration into a soil and subsequent storm runoff are highly dependent on the degree of saturation in the soil. Almost all major flood events are heavily influenced by the amount of rainfall that has occurred prior to the actual flood-causing rainfall.

Deforestation

The effect of trees on runoff has already been described, particularly with respect to water resources. There is also considerable evidence that a large vegetation cover, such as forest, decreases the severity of flooding. There are several reasons for this. The first has already been described, in that trees provide an intercepting layer for rainfall and therefore slow down the rate at which the water reaches the surface. This will lessen the amount of rainfall available for soil moisture and therefore the antecedent soil moisture may be lower under forest than for an adjoining pasture (NB this is not always the case, it is dependent on the time of year). The second factor is that forests often have a high organic matter in the upper soil layers which, as any gardener will tell you, is able to absorb more water. Again, this lessens the amount of overland flow, although it may increase the amount of throughflow. Finally, the infiltration rates under forest soils are often higher, leading again to less saturation excess overland flow.

The removal of forests from a catchment area will increase the propensity for a river to flood and also increase the severity of a flood event. Conversely the planting of forests on a catchment area will decrease the frequency and magnitude of flood events. Fahey and Jackson (1997) show that after conversion of native tussock grassland to exotic pine plantations a catchment in New Zealand showed a decrease in the mean flood peaks of 55–65 per cent. Although data of this type look alarming they are almost always taken from measurements at the small research catchment scale. At the larger scale the influence of deforestation is much harder to detect (see Chapter 11).

Urbanisation

Urban areas have a greater extent of impervious surfaces than in most natural landforms. Consequently the amount of infiltration excess (Hortonian) overland flow is high. In addition to this, urban areas are often designed to have a rapid drainage system, taking the overland flow away from its source. This network of gutters and drains frequently leads directly to a river drainage system, delivering more flood water in a faster time. Where extensive urbanisation of a catchment occurs, flood frequency and magnitude increases. Cherkauer (1975) shows a massive increase in flood magnitude for an urban catchment in Wisconsin, USA when compared to a similar rural catchment (see pp. 252–253). Urbanisation is another influence on flooding that is most noticeable at the small scale. This is mostly because the actual percentage area covered by impermeable urban areas in a larger river catchment is still very small in relation to the amount of permeable non-modified surfaces.

River channel alterations

Geomorphologists traditionally view a natural river channel as being in equilibrium with the river flowing within it. This does not mean that a natural river channel never floods, but rather that the channel has adjusted in shape in response to the normal discharge expected to flow through it. When the river channel is altered in some way it can have a detrimental effect on the flood characteristics for the river. In particular, **channelisation** using rigid structures can increase flood risk. Ironically, channelisation is often carried out to lessen flood risk in a particular area. This is frequently achieved, but in doing so water is passed on downstream at a faster rate than normal, increasing the flood risk further downstream. If there is a natural floodplain further downstream this may not be a problem, but if there is not, downstream riparian zones will be at greater risk.

Land drainage

It is common practice in many regions of the world to increase agricultural production through the drainage of ‘swamp’ areas. During the seventeenth and eighteenth centuries huge areas of the fenlands of East Anglia in England were drained and now are highly productive cereal and horticultural areas. The drainage of these regions provides for rapid removal of any surplus water, i.e. not needed by plants. Drained land will be drier than might be expected naturally, and therefore less storm runoff might be assumed. This is true in small rainfall events but the rapid removal of water through subsurface and surface drainage leads to flood peaks in the river drainage system where normally the water would have been slower to leave the land surface. So, although the drainage of land leads to an overall drying out of the affected area it can also lead to increased flooding through rapid drainage. Again, this is scale-related, as described further in Chapter 11.

Climate change

In recent years any flooding event has led to a clamour of calls to explain the event in terms of climatic change. This is not easy to do as climate is naturally so variable. What can be said though is that river channels slowly adjust to changes in flow regime which may in turn be influenced by changes in climate. Many studies have suggested that future climate change will involve greater extremes of weather (IPCC 2007), including more high-intensity rainfall events. This is likely to lead to an increase in flooding, particularly while a channel adjusts to the differing flow regime (if it is allowed to).

Case study

MOZAMBIQUE FLOODS OF 2000

During the early months of 2000, world news was dominated by the catastrophic flooding that occurred in southern Africa and Mozambique in particular. The most poignant image from this time was the rescuing of a young mother, Sophia Pedro, with her baby Rosita, born up a tree while they sought refuge from the flood waters. The international media coverage of the devastating flood damage and the rescue operation that followed has ensured that this flood will be remembered for a long time to come. It has given people the world over a reminder that flooding is a hydrological hazard capable of spreading devastation on a huge scale.

The floods of Mozambique were caused by four storms in succession from January through to March 2000. The first 3 months of the year are the rainy season (or monsoon) for south-eastern Africa and it is usual for flooding to occur, although not to the scale witnessed in 2000. The monsoon started early in southern Mozambique; the rainfall in Maputo was 70 per cent above normal for October–November 1999. This meant that any heavy rainfall later in the rainy season would be more likely to cause a flood.

The first flood occurred during January 2000 when the Incomáti and Maputo rivers (see Figure 7.12) both burst their banks, causing widespread disruption. The second flood occurred in early February, as the waters started to recede, except that now Cyclone Connie brought record rainfall to southern Mozambique and northern South Africa. The Limpopo river was as high as ever recorded (the previous high was in 1977) and major communication lines were cut. The third flood, 21 February until the end of February, occurred when Cyclone Eline moved inland giving record rainfall in Zimbabwe and northern South Africa, causing record-breaking floods. The Limpopo was 3 m higher than any recorded flood and for the first time in recorded history the Limpopo and Incomáti rivers joined together in a huge inundation. The extent of the flooding can be seen in the satellite images (see Figures 7.13 and 7.14). The fourth flood was similar in size to the second and occurred following Cyclone Glória in early March (Christie and Hanlon, 2001).

There is no doubt that the Mozambique floods were large and catastrophic. How large they are, in terms of return periods or average recurrence

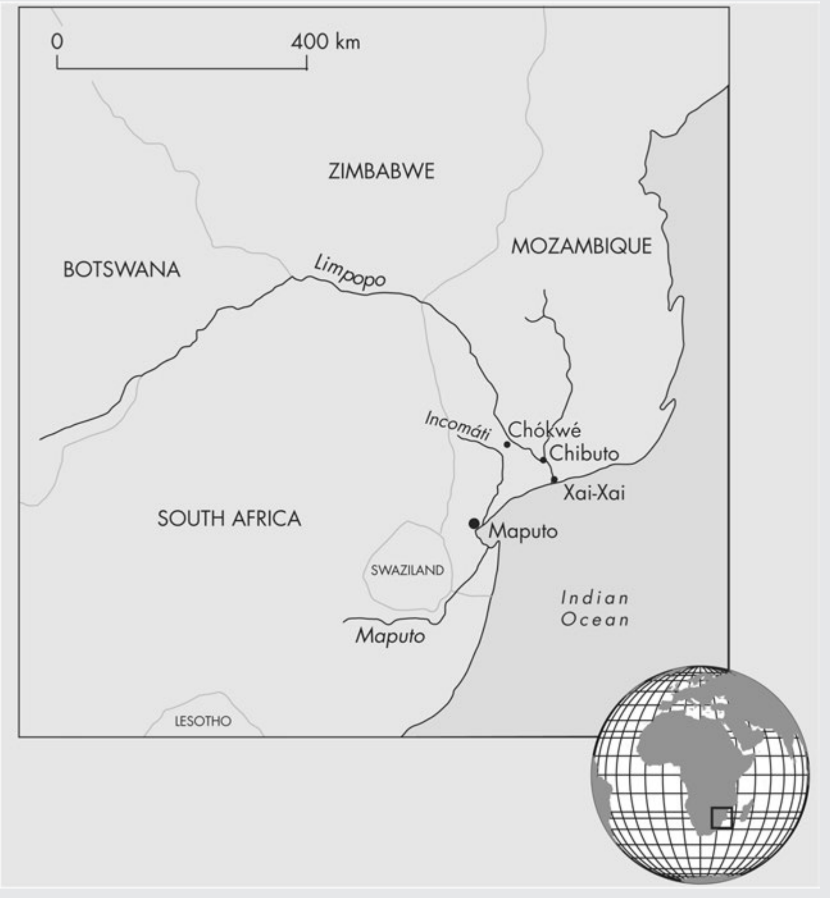


Figure 7.12 Location of the Incomáti, Limpopo and Maputo rivers in southern Africa.

Source: Image courtesy of G. Robert Brakenridge at the NASA-supported Dartmouth Flood Observatory

intervals (see Chapter 9) is difficult to assess. The major difficulty is to do with paucity of streamflow records and problems with measuring flows during flood events. On the Incomáti river the flow records go back to 1937, and this was the largest flood recorded. For the Limpopo there is some data back to the 1890s, and again this was the largest recorded flow event. On the Maputo river to the south the flood levels were slightly lower than a 1984 event. The difficulties in measuring riverflow during large flood events are well illustrated by the failure of many gauging stations to function properly, either through complete inundation or being washed away. Christie and Hanlon (2001) quote an estimate of the flood on the Limpopo having a 100-year average recurrence interval, although this is difficult to verify as most gauges failed. Smithers et al. (2001) quote an unpublished report by Van Bladeren and Van der Spuy (2000) suggesting that upstream tributaries of the Incomáti river exceeded the 100-year return period. Smithers et al. (2001) provide

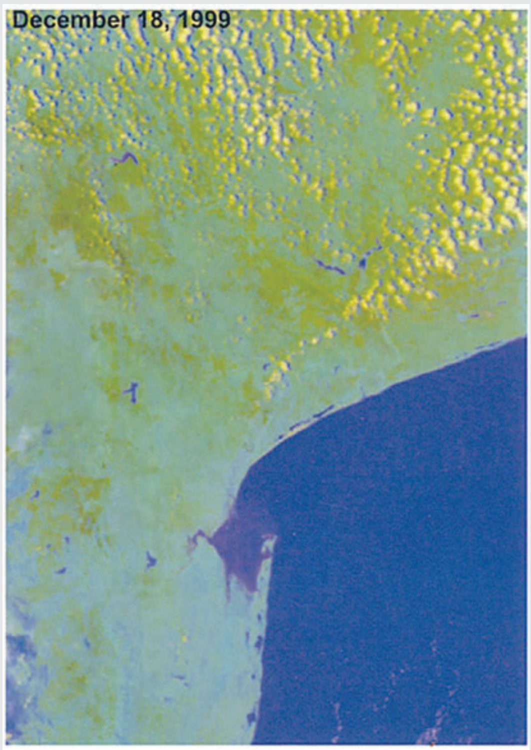


Figure 7.13 Satellite image of southern Mozambique prior to the flooding of 2000 (note location from Figure 7.12).

Source: Image courtesy of G. Robert Brakenridge at the NASA-supported Dartmouth Flood Observatory

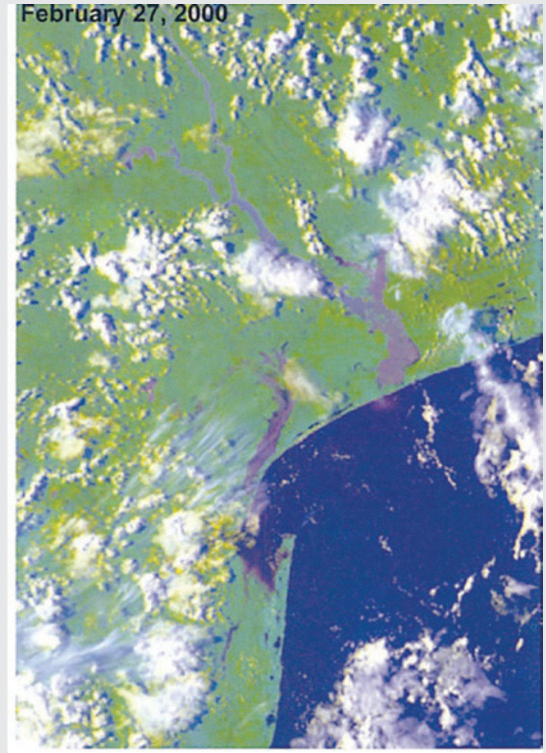


Figure 7.14 Satellite image of southern Mozambique following Cyclone Eline. The extensive flooding on the Incomáti, and Limpopo (top right of image) can be seen clearly.

an analysis of the 1–7-day rainfall for the Sabie catchment (a tributary of the Incomáti) which shows that in places the 200-year return period was exceeded. (NB this is an analysis of rainfall records not riverflow.)

The reasons for the flooding were simple, as they are in most cases: there was too much rainfall for the river systems to cope with the resultant stormflow. The river catchments were extremely wet (i.e. high antecedent soil moisture values) prior to the extreme rainfall, due to a prolonged and wet monsoon. One possible explanation for the severity of the rainfall is linked in with the ENSO (El Niño: Southern Oscillation) ocean–weather patterns in the Pacific. Christie and Hanlon (2001) present evidence that during a La Niña event (extreme cold temperatures in the western Pacific Ocean) it is common to see higher rainfall totals in Mozambique. However, this is not a strong relationship and certainly could not be used to make predictions. Figure 7.15 shows the monsoon rainfall at Maputo (averaged over two rainy seasons) and associated La Niña events. There may be some link here but it is not immediately obvious, particularly when you consider 1965–66 which had high rainfall despite it being

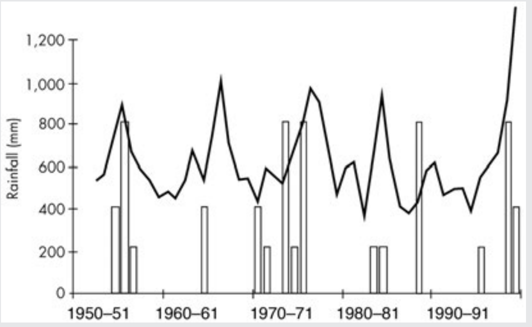


Figure 7.15 Rainfall totals during the rainy season (smoothed with a 2-year average) at Maputo airport, with vertical bars indicating the strength of La Niña events (on a scale of three: strong, medium, weak).

Sources: Rainfall data from Christie and Hanlon (2001); La Niña strength from NOAA

an El Niño event (often associated with drought in southern Africa).

What was unusual about the 2000 floods was that the tropical Eline cyclone (called typhoons or hurricanes elsewhere) moved inland, taking extremely high rainfall to Zimbabwe and northern South Africa. This is not normal behaviour for this type of storm and in so doing it created large floods in the headwaters of rivers draining into Mozambique. Flood warnings were issued by Zimbabwe and South Africa but the poor state of communications in Mozambique (exacerbated by the previous floods cutting communication lines) meant that they were not available to warn people on the ground. In all, 700 people died as a result of the floods and 45,000 people were displaced. It is estimated that it will cost US$450 million to repair damage to the infrastructure in Mozambique (Christie and Hanlon 2001). This is not the total cost of the flood, which is far higher when loss of income and loss of private property are included. These costs will never be fully known as in many lesser-developed countries the costs are borne by individuals without any form of insurance cover.

In many ways there are no new lessons to learn from the Mozambique floods of 2000. It is well known that adequate warning systems are needed (but expensive) and that people should be restricted from living in flood-prone areas; but this is difficult to achieve in a poor country such as Mozambique. The cause of the flood was a huge amount of rainfall and the severity was influenced by the antecedent wetness of the ground due to a very wet monsoon.

Case study

BRISBANE FLOODING IN 2010–11

The case study above of flooding in Mozambique in 2000, was largely driven by the La Niña weather system. In 2017, conditions were repeated, and another major flood occurred in Mozambique, showing the significant influence of this large-scale climatic phenomenon. Similarly, a strong La Niña event during the southern hemisphere summer of 2010–11 was a contributor to the high rainfall totals, mainly through extra warm sea surface temperatures off the Queensland coast. As a consequence, large floods occurred in southern Queensland, Australia during the summer of 2010–11.

Floods are often classified based on probability (e.g. the average recurrence interval), observed maxima (e.g. the largest flows in a river since records began) or the adverse effects (e.g. number of people affected or the damage costs). On all three of these criteria, the Brisbane floods of early 2011 were very large events. The rainfall in the

Brisbane River catchment had an annual exceedance probability of 1 in 150 but in parts of the catchment it exceeded 1 in 500 (Nathan 2012) (Figure 7.16).

In Brisbane City although it was the seventh highest recording of river height since records began in 1840 and the highest since 1974, an analysis of water volumes suggests that it was nearly double the size of the 1974 flood. In terms of adverse effects, 37 people were killed during the floods; 29,000 homes or businesses were inundated and it is estimated the cost of insurance and recovery work was $5 billion Australian dollars (approximately $3.9 billion US dollars).

It is interesting to compare the size and impact of the 2011 and 1974 floods as it crosses between physical hydrology (i.e. hydrological processes) and water resource management, but first it is necessary to give some geographic context to the floods. The floods are referred to as the Brisbane floods but they affected a much wider area of southern Queensland. The Brisbane River catchment is 13,570 km^2 and although the majority of land is used for pastoral grazing and forestry it also includes the cities of Brisbane (2011 population of 1.98 million people) and Ipswich (200,000 people) in its lower reaches.

The peak of the flood in Brisbane City was on 13 January, 2011 but it was the culmination of extreme rainfall from late November 2010. Six large rainfall events during this time led to December 2010 being the wettest December

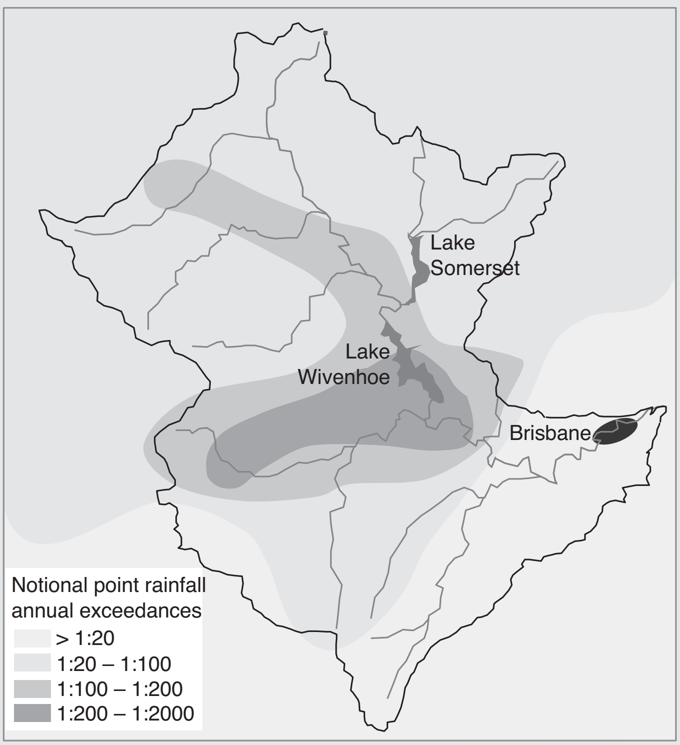


Figure 7.16 The Brisbane catchment on the east coast of Australia showing an interpolated distribution of annual exceedance probabilities for rainfall (years).

Source: Nathan (2012), used with permission on record for southern Queensland and the soil moisture being fully saturated; perfect conditions for a flood to occur if another large rainfall event comes.

Within the Brisbane catchment there are two major pieces of water infrastructure: the Wivenhoe (finished 1958) and Somerset (finished 1985) Dams. Although the primary purpose of both dams and the reservoirs behind them is to supply potable water to Brisbane, Ipswich and other nearby towns, they also provide hydro-electric power, recreational opportunities and most particularly for the Wivenhoe Dam, flood mitigation potential. The fact that the 2011 flood gave a lower recorded river height in Brisbane despite being a much larger volume of water than the 1974 flood is partly due to the flood mitigation provided by the Wivenhoe dam. However, there is some controversy around this as many people blame the Wivenhoe Dam operation for exacerbating the flood event.

The spatially averaged rainfall total for the Brisbane catchment for 10–12 January, 2011 was 286 mm; less than the 349 mm 3-day total for 25–27 January, 1974 (van den Honert and McAneney 2011). However, the inflows at gauging stations above the Wivenhoe Dam were the highest on record (Nathan 2012), which suggests that the role of antecedent soil moisture was important in making for a larger flood volume than in 1974. Data from Seqwater, the company operating the Wivenhoe and Somerset Dams, show that on 7 January, both dams were above full storage capacity and consequently they released extra water into the rivers. By 11 January they had enough storage available that they could release only 60 per cent of the inflows (i.e. 40 per cent of the inflow was stored) and subsequently it is estimated the flood peak in Brisbane City was reduced by 2 m and a further 14,000 properties were saved from inundation (van den Honert and McAneney 2011).

Flood mitigation is an important water resource function but it is often much easier in theory than practice. During a large flood event such as the Brisbane Floods of 2011 the dam operator is balancing holding as much water back as possible while retaining storage potential for predicted rainfall events. The accuracy of rainfall prediction becomes paramount despite rainfall being a notoriously difficult process to predict precisely. The independent hydrologist engaged by the Queensland Commission of Inquiry to investigate the role of Wivenhoe Dam operation concluded that the dam operations had achieved the best possible flood mitigation effect.

At the time of the Brisbane Floods there was intense media speculation on the causes of the flooding with much speculation on the role of the Wivenhoe Dam operations. The vindication provided by the Commission of Inquiry is gratifying for the dam operators but it is worth considering why the media, as reflectors of society, were so intense in their scrutiny. There are several factors in this. The first is that the summer of 2010–11 was preceded by an extreme dry period, so much so that the year before there were serious concerns that Brisbane would run out of potable water; therefore the prevailing narrative was of low water levels in the dams above Brisbane. A second point is that there had been no major floods since 1974 so that flooding was not in the collective memory of many Brisbane residents. In addition to that the Wivenhoe Dam had been built since 1974 which no doubt gave a sense of false security that flooding was no longer an issue for Brisbane. A final point is that our twenty-first-century society likes to have an instant scapegoat for natural disasters like floods. Nathan (2012) uses the 2011 Brisbane floods to draw 'lessons from large floods'. Amongst the lessons learned are the following: big rains cause big floods; communicating risk is difficult; and hydrologic complexity cannot be ignored. These lessons apply across the world and are worth water resource managers and hydrologists having pinned on their office walls!

RUNOFF IN THE CONTEXT OF WATER QUALITY

The route that water takes between falling as precipitation and reaching a stream has a large influence on water quality. The nutrient level of water is heavily influenced by the length of time water spends in contact with soil. Water that moves quickly into a river (e.g. overland flow) is likely to have a lower nutrient level than water that moves slowly through the soil as throughflow and/or groundwater. However, water that has travelled as overland flow may have a higher level of suspended solids picked up from the surface, so it may appear less pure.

In considering issues of land-use change and water quality, an important consideration is the time taken for water to reach the stream. It is important to realise that when groundwater responds to a rainfall event by emitting water into a stream, it is frequently operating as a pressure wave response to rainfall recharge, i.e. the water entering the stream is not the same water that infiltrates and causes the response. This means that water entering the stream may be several years (or more) older and unaffected by the current land use change.

ESSAY QUESTIONS

1. **Explain and compare two contrasting theories of surface runoff generation.**
2. **Consider two catchments of the same area, general topography and land cover. The one catchment is characterised by predominantly sandy soils whilst the other is a clay catchment. Evaluate the likely runoff generation mechanisms in each catchment with particular reference to stormflow generation theories. Illustrate your answer with representative hydrographs.**
3. **Critically evaluate the statement that infiltration excess overland flow is the only type of overland flow relevant to stormwater runoff estimation.**
4. **Understanding of runoff processes has increased considerably since the 1930s. Discuss.**
5. **Give an account of the principal mechanisms of runoff generation and critically evaluate the statement that infiltration excess overland flow is the only type of overland flow relevant to flooding.**

FURTHER READING

Anderson, M.G. and Burt, T.P. (eds) (1990) *Process studies in hillslope hydrology*. Wiley, Chichester.
A slightly more modern update on Kirkby (1978).

Bevan, K. (2006) *Streamflow generation processes*. IAHS Benchmark Papers in Hydrology. International Association of Hydrological Sciences. IAHS Press, Wallingford.
Part of the excellent IAHS series collating key original papers with added commentary. This volume takes a chronological perspective showing the evolution of theory.

Kirkby, M.J. (ed.) (1978) *Hillslope hydrology*. J. Wiley & Sons, Chichester.
A classic text on hillslope processes, particularly runoff.

Parsons, A.J. and Abrahams, A.D. (1992) *Overland flow: Hydraulics and erosion mechanics*. UCL Press, London.
An advanced edited book; good detail on arid regions.

Smith, K. and Ward, R.C. (1998) *Floods: Physical processes and human impacts*. Wiley, Chichester. A text on flooding.

Wohl, E.E. (2000) *Inland flood hazards: Human, aquatic and riparian communities*. Cambridge University Press, Cambridge.
A text on flooding with many case studies.

8

MEASURING CHANNEL FLOW

LEARNING OBJECTIVES

When you have finished reading this chapter you should have:

- A knowledge of the techniques and instruments used for measuring streamflow directly.
- An understanding of the challenges involved in maintaining an accurate flow record.
- A knowledge of techniques used to estimate streamflow.
- How to construct a stage vs discharge relationship (a rating curve).

Being able to measure the volume of water flowing down a stream in a set amount of time is a fundamental task for any hydrologist. The techniques and research into the measurement of streamflow are referred to as **hydrometry**. Hydrometry is the area of hydrology that has probably seen the greatest changes in the past 10–15 years, driven by an integration of advanced electronic instrumentation into environmental science. However, the fundamental principles of streamflow measurement remain the same. Streamflow measurement can be subdivided into two important subsections: *instantaneous* and *continuous* techniques.

INSTANTANEOUS STREAMFLOW MEASUREMENT

Velocity–area method

Streamflow or discharge is a volume of water per unit of time. The standard units for measurement of discharge are m^3/s (cubic metres per second or *cumecs*). If we rewrite the units of discharge we can think of them as a water velocity (m/s) passing through a cross-sectional area (m^2). Therefore:

$$m^3/s = m/s \times m^2 \tag{8.1}$$

The **velocity–area method** measures the stream velocity (m/s) and the stream cross-sectional area (m^2) and then multiplies the two together. In practice this is carried out by dividing the stream into small sections and measuring the velocity of flow going through each cross-sectional area and applying Equation 8.2.

$$Q = v_1 a_1 + v_2 a_2 + \dots v_i a_i \tag{8.2}$$

Where Q is the streamflow or discharge (m^3/s), v is the velocity measured in each trapezoidal cross-sectional area (see Figure 8.1), and a is the area of the trapezoid (usually estimated as the average of two depths divided by the width between).

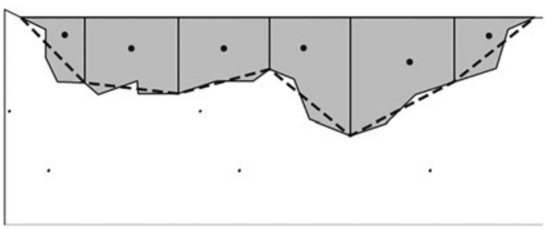


Figure 8.1 The velocity–area method of streamflow measurement. The black circles indicate the position of current meter velocity readings. Dashed lines represent the triangular or trapezoidal cross-sectional area through which the velocity is measured.

Figure 8.2 Flow gauging a small stream using a mechanical current meter.

Making a streamflow measurement

There is an international standard for the measurement of discharge in streams, rivers and other open channels (ISO 748:2007). This ISO standard sets out precise details for measurements to meet the highest standard; the methods described in this book are based on this standard but give broader guidelines.

The first consideration when making a streamflow measurement is to select a good reach of the stream. The channel should be straight and uniform in bed and slope down the reach (i.e. it doesn't move from a pool to a riffle, or vice versa, in the reach). Areas with reverse flow or back waters should be avoided and the cross section to be measured should be clear of overhanging trees reaching the water or other obstacles such as large rocks.

Once a suitable reach has been found, a tape measure or 'tag line' should be strung across the stream to measure width and establish position of verticals in the cross section. The number of verticals required in the cross-sectional areas that are used in a discharge measurement depend upon the width and smoothness of the stream bed. If the bed is particularly rough it is necessary to use more cross-sectional areas so that the estimates are as close to reality as possible (note the discrepancy between the dashed and solid lines in Figure 8.1). The ISO standard for number of verticals is set out in Table 8.1. Note for channels greater than 5 m in width you may have to do a trial discharge measurement and check whether you have enough verticals to fulfil the criteria of between 5–10 per cent of total flow.

The measurement of stream velocity is done using a current meter. The development of different types of current meters has been a huge advance in hydrometry in the past 20 years and different types of flow meters are described in the following section.

In the velocity–area method it is necessary to assume that the velocity measurement is representative of all the velocities throughout the

Table 8.1 ISO standard guidance on number of verticals required for accurate assessment of stream discharge

Channel width	*Number of verticals*
Less than 0.5 m	5–6
0.5 m to 1.0 m	6–7
1.0 m to 3.0 m	7–12
3.0 to 5.0 m	13–16
Greater than 5.0 m	Enough so that the discharge in each vertical is between 5 and 10% of the total

Note: The number of verticals does not include the two zero depth measurements at either side of the cross section.
Source: ISO 748:2007

cross-sectional area. When using a mechanical current meter it is not normally possible to take multiple measurements so an allowance has to be made for the fact that the water travels faster along the surface than nearer the stream bed. This difference in velocity is due to friction exerted on the water as it passes over the stream bed, slowing it down. The International Standard sets out methods for one, two, three, five and six depth profiles but the most common is to use one or two depths. For a single sample of velocity the current meter should be 60 per cent of the stream depth – that is, in a stream that is 1 m deep the sampling point should be 0.6 m below the surface or 0.4 m above the bed. In a deep river it is good practice to take two (or more) measurements. If using the two-depth sample method then one measurement should be at 20 per cent and the other at 80 per cent of total depth and two velocities are averaged.

When discharge is measured with a waded cross section, the current meter (whether mechanical, electromagnetic or ADCP) is attached to a graded metal pole, called a **wading rod**. Although relatively simple in design, a good wading rod is an essential tool for any field hydrologist. At the bottom of a rod is the baseplate which rests firmly on the stream bed. The rod has graduated markings up the pole so that the depth of the stream can be easily read. The best form of wading rod is called a top setting rod which, as well as having the depth adjustment for the meter above the water, also has an easy calculator for 60 per cent of the total depth. Although this may not be important on a lovely warm day it is very important on a cold winter's day when you do not want to immerse your hands in water for any longer than you have to. It is also surprising how difficult it is to make a mental calculation of 60 per cent of the total depth when you are very cold – hence the necessity of a good wading rod!

When taking the velocity measurement it is important that the current meter points directly into the stream flow. If a good reach of river has been chosen then the direction of flow should be at right angles to the tape strung across the stream to measure width. The International Standard sets out that each velocity measurement must be for a minimum of 30 seconds; the actual time taken per reading is instrument dependent and will be set out in the instructions for use.

Where there is no current meter available it may be possible to make a very rough estimate of stream velocity using a float in the stream and measuring the time it takes to cover a measured distance. When using this method, allowance must be made for the fact that the float is travelling on the surface of the stream at a faster rate than water closer to the stream bed and that the velocity will not be uniform across the whole profile. If the faster surface velocity is ignored then the estimated velocity will be higher than the overall average and streamflow over-estimated. Somewhat surprisingly given the approximate nature of this measurement technique, the International Standard does set out procedures for using floats in velocity measurement. ISO 748:2007 suggests that surface velocity should be multiplied by somewhere between 0.84 and 0.9 to give the actual average velocity through the profile. The higher values are for when there is a smooth stream bed and lower for rough beds (i.e. many boulders).

Figure 8.2 shows the measurement of streamflow in a river by wading across with the mechanical current meter held in front and pointing upstream. This is non-continuous; i.e. the person stops at points across the profile and measures the velocity for a period of time, usually 30 seconds or more. They then work their way across the stream to create the type of profile shown in Figure 8.1. In a much larger river, or a river in flood, it is not possible to wade across so the velocity measurements have to be taken in another way. On a very large river, like the Mississippi, this would normally be done from a boat that slowly crosses the river with a current meter attached to the front. The boat needs to stop at set locations and remain stationary (i.e. drive against the current) while the velocity measurement is taken – a very skilful task. In large rivers it is normal to take velocities at a series of depths to gain an average. For rivers in flood it is frequently unsafe to operate a boat (e.g. Figure 8.3) so velocity measurements are taken from a nearby bridge or from a cage suspended above the river on a cableway (see Figure 8.4). For very large flows (e.g. during a flood) the propeller on the current meter needs to be very large and weighted heavily to ensure it sinks down through the river profile rather than be swept downstream or stay near the surface. When gauging from a bridge the measurement needs to be taken enough upstream from the bridge so that the flow around bridge supports does not interfere with the velocity measurement. This normally means some form of crane rig that suspends the current meter several metres upstream from the bridge. Needless to say, whether from a bridge or a cableway, flood gauging is a very dangerous task as the river velocities are likely to be high and flood debris can catch on the current meter, dragging it downstream.

Figure 8.3 A river in heavy flood. Measuring the flow here could be done from the nearby bridge but it is a very dangerous job and the velocities will be influenced by the turbulent nature of the river and increased flow velocities around the bridge supports.

Figure 8.4 Cableway used for gauging in a large river. The current meter is suspended from the cable at set points across the river and a velocity profile is measured.

The velocity–area method is an effective technique for measuring streamflow, but its reliability is heavily dependent on the sampling strategy. The technique is also less reliable in small, turbid streams with a rough bed (e.g. mountain streams). Under these circumstances other streamflow estimation techniques such as **dilution gauging** may be more applicable (see streamflow estimation section).

Types of current meter

Mechanical current meters

A mechanical flow meter is a form of propeller inserted into the stream which records the number of propeller revolutions with time as the water flows in and around the meter. The number of revolutions in a set time is easily converted into stream velocity using the calibration equation supplied with the current meter. There are many different sizes and types of mechanical current meters, all of which are designed for different stream sizes and velocities. It is important that the correct meter is used in the right conditions otherwise inaccurate velocities will be measured. Large propellers are for use in large swift rivers (or small rivers in flood) but will be inappropriate for small rivers at low flow as the propellers will be too big for stream depth. In a vice versa form small propellers are not able to rotate quickly enough in high velocity streams and underestimate the stream velocity.

Up until around the turn of the twenty-first century, stream velocity was most commonly measured with mechanical current meters. A good mechanical current meter is a robust and reliable tool for measuring instantaneous streamflow but it is gradually being replaced by more advanced technology using Electromagnetic or Acoustic Doppler Current Profiling (ADCP) techniques.

Electromagnetic current meters

Electromagentic current meters are widely used for measuring the velocity of fluids in pipes but a variation of the technology can be used to measure water velocity in an open channel. The fundamental theory underpinning electromagnetic flow meters is that of electromagentic induction; first discovered by Michael Faraday in 1831. Faraday's Law describes how a voltage will be induced in a conductor moving through a magnetic field. In the case of streamflow measurement the conductor is the body of water flowing past a magenetic field between two magnets. When the length of conductor (i.e length of water measured over) and the strength of magnetic field remain fixed the magnitude of the induced voltage is directly proportional to the velocity of the water. One of the earliest electromagentic flow meters was built by Faraday himself in 1832 when he unsucessfully tried to measure the discharge down the River Thames in London as it flowed under Waterloo Bridge. The biggest problem Faraday faced was that he could not measure the voltage accurately enough, something that has now been overcome.

Electromagnetic current meters have an advantage over mechanical meters in that they don't have propellers interfering with flow and don't easily get fouled up with debris. They are expensive pieces of equipment and generally have not had as high an uptake as ADCP technology.

Ultrasonic flow measurement

Ultrasonic flow measurement relies on precise measurement of the difference in transit time of ultrasonic pulses propagating with and against flow direction. The most common form of ultrasonic flow measurement uses Acoustic Doppler Current Profiling (ADCP, also called Acoustic Doppler Velocimetry – ADV). The growth of ADCP technology has led to huge advances in hydrometry over the past 20 years. ADCP allows us to measure stream velocity in an integrated manner across a river; i.e. it measures velocity at a series of depths in the profile all at once rather than at a single point like a mechanical current meter. To understand ADCP you need to have a basic understanding of the Doppler Effect; named after Austrian physicist Christian Doppler who first proposed the idea in 1842. A simple analogy to demonstrate the Doppler Effect is if you stand on the side of a road while a car drives past you. As the car goes past you the engine noise appears to change from a higher to a lower sound while for the driver the engine noise stays the same. As the car approaches, each sound wave is emitted from a position closer to you than the previous wave. Therefore each wave takes slightly less time to reach the observer than the previous wave. The opposite happens as the car moves away from you; effectively lengthening the sound wave. For the car driver it always sounds the same because the sound waves are always coming from the same distance away. To complicate things further, the Doppler Effect will sound more pronounced the faster the car is going and will also depend on the wind speed. The Doppler Effect may therefore result from motion of the source (the car), motion of the observer (you), or motion of the medium (air).

In order to measure river velocity, the Doppler Effect is utilised by an instrument emitting an ultrasonic acoustic burst, which then measures the change in wavelengths (frequency) of reflections from particles moving with the flow. The faster the movement of particles in the stream, the greater is the distortion in wavelength that the ADCP sensor detects. An ADCP instrument can measure in real time a complete velocity profile for a stream containing particles in suspension such as sediment or air bubbles. Sound travels at approximately 1,500 m/s in water (dependent on water purity and depth) so the instrumentation used in this type of flow gauging needs to be extremely precise and be able to measure in nanoseconds. It is only with advances in electronics since the 1990s that a portable field instrument has been able to be developed; i.e. sending high frequency ultrasonic waves and then being able to measure the frequency of returning waves requires sensors that weren't available 30–40 years ago and a computing power that can now be found in simple silicon chips. Early ADCP instruments required a significant amount of suspended sediment or air bubbles in the water to operate well; a problem for very clear streams. More modern instruments are less prone to this as they can detect very small air bubbles that exist in almost all streams.

There are many different types of ADCP instruments all of use in different streams and rivers. There is a small pole-mounted ADCP such as the SonTek Flow Tracker (see Figure 8.5) that is used in the same way as a mechanical current meter with the user wading across a stream and taking a velocity profile at points across the stream. An advantage of the Flow-Tracker type instrument over a mechanical current meter is that velocity is measured in a three dimensional cone in front of the instrument, therefore providing an average velocity over a greater area than the single point from a mechanical meter. In general, the Flow Tracker can be used in a wider range of conditions (e.g. shallower water and lower velocity) than a mechanical meter.

Figure 8.5 Portable ADCP (SonTek Flow Tracker) being used to measure a small stream.

There are also larger ADCP instruments that can be mounted on boats to provide a velocity profile across a stream (see Figure 8.6). Recent models can be mounted on a remote controlled small boat that can be driven across the river while the user stands on the bank; thus improving the health and safety for the field hydrologist. For a boat mounted ADCP, sound waves reflecting off the river bottom can be used to define the depth profile plus the velocity (i.e. both speed and direction) of the boat itself; a feature referred to as 'bottom tracking'. Using this feature, a full profile of velocity and depth can be derived which automatically integrates the velocity and area in a single measurement of flow.

A problem can arise with bottom tracking when the bed of the river is moving; for instance during a flood event the sand and gravel on a river bed can be moving with the faster velocity of the river. In this case the bottom tracking doesn't work and a Geographic Positioning System (GPS) unit is required on the ADCP to keep track of position across the river.

Figure 8.6 A boat mounted ADCP carrying out a cross section. The disc at the bottom of the pole on the near side of the boat (just touching the water surface) is the ADCP sensor.

CONTINUOUS STREAMFLOW MEASUREMENT

The methods of instantaneous streamflow measurement described above only allow a single measurement to be taken at a location. Although this can be repeated at a future date it requires a continuous measurement technique to give the data for a hydrograph. There are three different techniques that can be used for this: stage–discharge relationships, flumes and weirs, and ultrasonic flow gauging. There are two International Standards that cover continuous stream flow measurement (ISO 1100–1 and 1100–2).

Stage vs discharge relationship

River **stage** is another term for the water level or height. Where multiple discharge measurements have been taken (i.e. repeat measurements using velocity–area method) it is possible to draw a relationship between river stage and discharge: the so-called **rating curve**. An example of a rating curve is shown in Figure 8.8. This has the advantage of allowing continuous measurement of river stage (a relatively simple task) that can then be equated to the actual discharge. The stage–discharge relationship is derived through a series of velocity–area measurements at a particular site while at the same

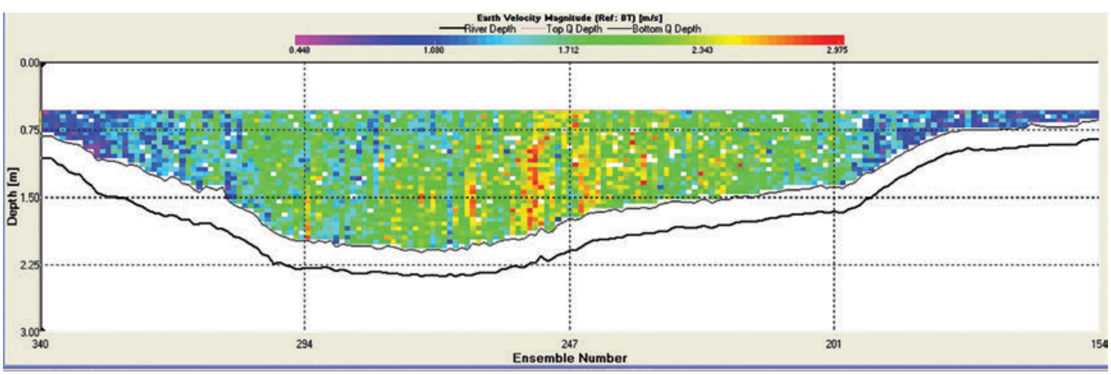


Figure 8.7 Computer output from the ADCP profile in Figure 8.6. Colour shows the velocity at depths (vertical axis). The red colour in the middle at mid-depth is around 2.5 m/s; the blue and purple in the shallow edges is around 0.78 m/s.

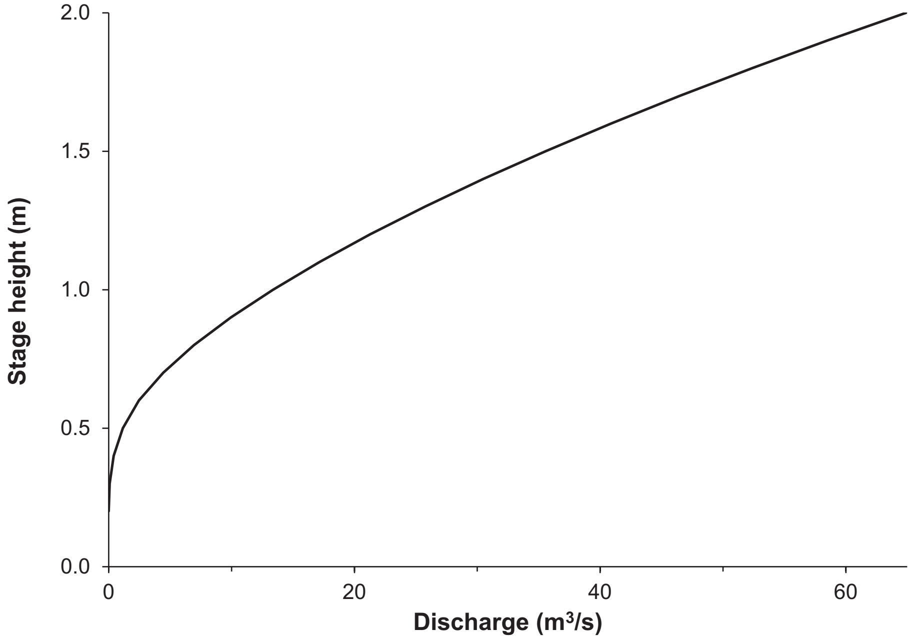


Figure 8.8 A rating curve for the river North Esk in Scotland based on stage (height) and discharge measurements over a 27-year period.

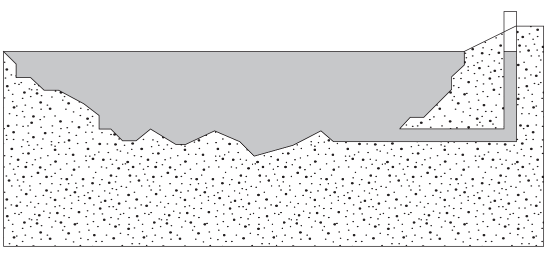


Figure 8.9 Stilling well to provide a continuous measurement of river stage (height). The height of water is measured in the well immediately adjacent to the river (see photograph in Figure 8.10 below).

time recording the stage height (see Figure 8.9). As can be seen in Figure 8.8, the rating curve is non-linear, a reflection of the river bank profile. When the river is low it is confined to a small area so that a small change in discharge (i.e. increased flow) will cause a significant rise in stage height. As the river fills up between banks it takes a greater volume of water to cause the same change in stage height than at low levels.

An accurate stage vs discharge relationship is dependent on frequent and accurate measurement of river discharge, and a static river bed profile. If the river bed profile changes, the stage vs discharge relationship (rating curve) will change and the historic relationship will no longer be valid. The three most common reasons for a change in bed profile are the bed of a river being scoured out during a flood event; the bed being raised up from sediment deposition during a flood event; or aquatic plants growing in a stream causing the velocity to slow down and the water level to rise (see Case Study). In a gravel-bed river (e.g. Figure 8.10), bed movement is a constant issue which may mean adjustments to the rating curve after every fresh or flood. In Figure 8.12 the record of stage seems to have shifted upwards following the large flood in the middle of the record. Although it is possible that the flow remained higher following the flood, the fact that it never drops to anywhere near the pre-flood stage make it more likely the bed has aggraded (built up in level). Flow gaugings during and after the

Figure 8.10 A hydrometric station with the stilling well beside a gravel bed river.

Figure 8.11 A stilling well in a large concrete tower beside a mountain river. In this case the tower has to be high to record high levels in large floods. Note the external staff gauge attached to the concrete tower.

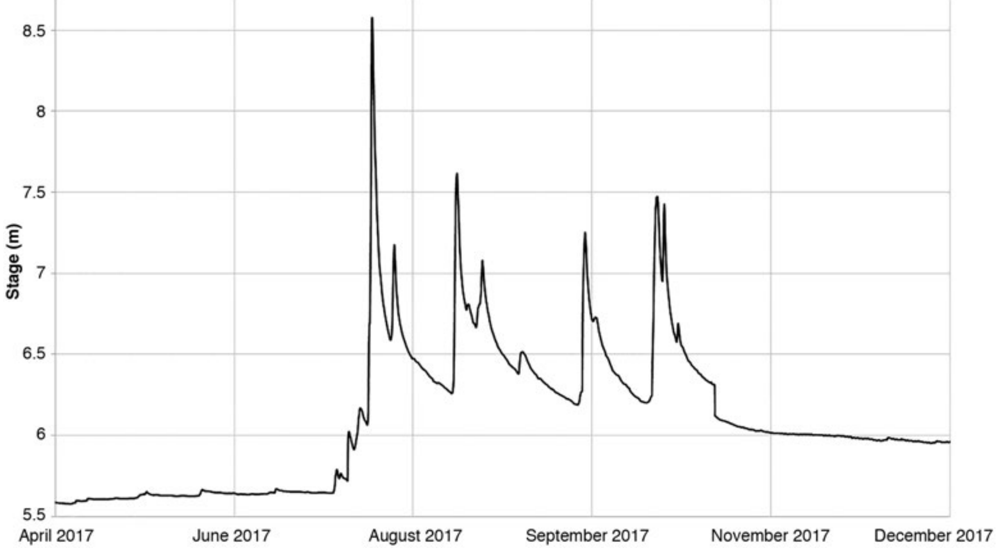


Figure 8.12 Stage record for a gravel bed river (Selwyn River) with clear evidence of bed aggradation after the large flood event in the middle of the record. The measured flows in July and late November 2017 were the same (830 l/s), pointing to bed aggradation causing the shift in base level.

flood confirmed that the stage vs discharge relationship had changed and a new rating curve was required. If a new rating curve wasn't used then the flow would have been over-estimated following the flood. The assumption of a static river bed profile can sometimes be so problematic that there is a need for a concrete structure (e.g. flume or weir) to be installed to maintain stability.

Case study

MACROPHYTE GROWTH AFFECTING A RATING CURVE

The Halswell River is a small spring-fed river in the South Island of New Zealand (Figure 8.13). The river shows all the classic characteristics of a groundwater fed system: it continues flowing steadily through the summer and does not have a huge range of flows – it doesn't get very low, nor does it get very high except in exceptionally large storm events. Because of the catchment intensive agricultural land use the water is nutrient rich; this combined with a lack of frequent flushing flows makes the river an ideal growing place for aquatic plants (macrophytes). As the plants grow they choke the stream and lessen the cross sectional area. This means that either the velocity has to increase or the water level rises to accommodate the smaller cross sectional area.

Figure 8.13 Macrophyte growth in the Halswell River constricting flow and both raising the water level and increasing velocity.

In a low gradient river like the Halswell, the choking of the stream can happen over a long distance and it is not feasible to continuously clear the stream of aquatic plants. During the Southern Hemisphere late summer of February to March 2013 there was a continuous rise in the water level which at first consideration would suggest an increase in flow (see the rise on the solid line in Figure 8.14). However, regular gauging of the stream showed that each time the velocity had decreased, the flow had actually decreased (see Figure 8.14). The macrophytes were changing the cross sectional area as they grew. To maintain an accurate flow record, it was necessary to continually adjust the rating curve. The difficulty with this is that it is a gradual adjustment as the plants grow; as opposed to a change from a discrete event like a flood. As a result the continuous measurement of low flow in nutrient-rich, spring-fed streams like the Halswell River is a challenge for field hydrologists.

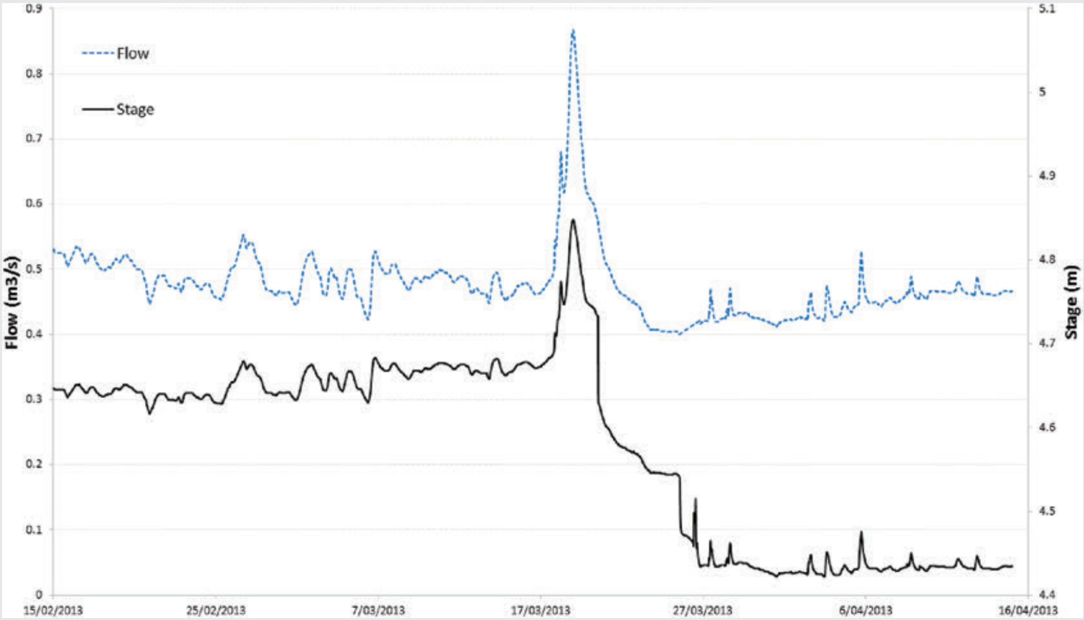


Figure 8.14 Stage (solid line) and flow (dashed) record for the Halswell River. The drop in stage after 18 March, 2013 is due to macrophyte clearance downstream of the flow recorder. Note the flow does not drop correspondingly.

One of the difficulties with the stage–discharge relationship is that the requirement of frequent measurements of river discharge lead to many measurements taken during periods of low and medium flow but very few during flood events. This is for the double reason that: floods are infrequent and unlikely to be measured under a regular monitoring programme; and the danger of streamflow gauging during a flood event. The lack of data at the extreme end of the stage vs discharge curve may lead to difficulties in interpreting data during peak flows. The error involved in estimating peak discharge from a measured stage vs discharge relationship will be much higher at the high flow end of the curve. However once the top of a rating curve is established (i.e. flow measurement carried out during a peak flood) that point is less likely to be affected by bed changes as the depth will be great and a small change from bed scoring or deposition will cause less of an effect than when at low flow.

When interpreting data derived from the stage–discharge relationship it is important that the hydrologist bears in mind that it is stage height that is being measured continuously and from this stream discharge is estimated via a rating curve relationship (i.e. it is not a direct measurement of stream discharge).

Case study

HOW TO CONSTRUCT AND MAINTAIN RATING CURVES

The need to have good accurate rating curves underpins the work of any field hydrologist. Without them, flow data are unreliable and any further statistical analysis of streamflow data will be of dubious quality. However, there are no hard and fast rules for rating curve construction and it can be viewed as an art as much as science. An experienced field hydrologist will get 'the feel' for how a rating curve should look and know very quickly whether it needs changing. For the less experienced there are many tips that will be helpful.

A curious feature of rating curves is that by tradition discharge is on the x-axis (horizontal) with stage on the vertical y-axis when they should really be the other way around. In statistics convention has it that the independent variable is on the x-axis with the predicted or dependent variable on the y-axis. In a rating curve we are predicting discharge (dependent) from the stage height (independent); so by mathematical convention they should really be the other way around.

The absolute basics for constructing a rating curve is to connect a series of flow measurements (i.e. gaugings) with concurrent measurements of river stage (height). Flow is measured with a current meter using the velocity-area method described elsewhere in this chapter and stage is measured from a manual staff gauge at the flow recorder. To construct a rating curve from scratch you will be required to have flow measurements at a range of flows (low and high). It will normally take several months of flow gaugings at a site before there are enough gaugings to construct the rating curve. Even then it is highly unlikely that there will be gaugings across the full range of flows (see Figure 8.15).

Measured flows are plotted showing error bars (normally between 5–10 per cent depending on measurement method and flow volumes) whereas the stage measurement is normally accurate enough to not require an error bar to be drawn.

The 'art' of drawing a rating curve now comes in. Most hydrological computer programmes offer interpolation techniques for joining between gaugings but there will always be some interpretation in how to smooth between points. The most important point to note is that for all rivers in non-rectangular cross sections the rating curve will be non-linear. This is because as the

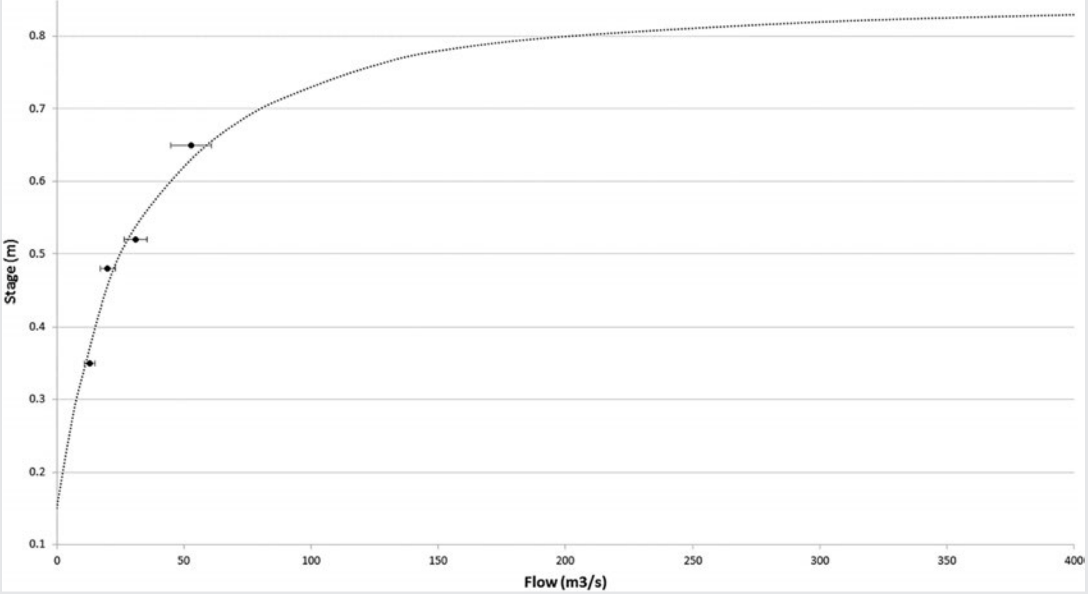


Figure 8.15 Starting to draw the rating curve. The flow gaugings are shown with error bars to reflect uncertainty in flow measurement with the veocity–area method.

river level rises the width increases (i.e. the river spreads out) and the cross sectional area changes in a non-linear fashion. The exact shape of the non-linear curve is dependent on the shape of the river channel. As can be seen in Figure 8.15 the curve will not normally go exactly through each gauging point but you would expect it to be within gauging error. If this is not the case then the gauging needs careful checking and you may need to reconsider whether the bed level has stayed static between gaugings.

A second important point is that ratings curves should be smooth, not having erratic dips and highs to accommodate gaugings. With time and regular gaugings at a site more points can be added to the top and bottom of the curve, thus getting a better idea of overall shape.

The effort required to maintain a rating curve is dependent on how stable the site is. For a well controlled site with a stable bed you would not expect the rating curve to change except in extreme flood events. For gravel-bed rivers the changes can be frequent and it is a challenging job to keep accurate flow records. When gaugings are plotted with error bars and these do not intersect the rating curve it is time to change the rating curve (see example in Figure 8.16). Although you can do this on the basis of one gauging it is normal (and wise) to have several gaugings so that the new shape can be established. If the rating curve is at a long-established site then the first thing to do is retrieve previous rating curves and see if any of these fit the gaugings. You should do this before starting to draw a new curve as rivers frequently re-establish old channel forms.

At an unstable site, generally you will see more change at the lower end of the rating curve than at the higher end. This is because bed degradation and aggradation has a much larger effect on low flows than on higher flows where the volume of water is so large. So generally different rating curves tend to converge on the same, or similar

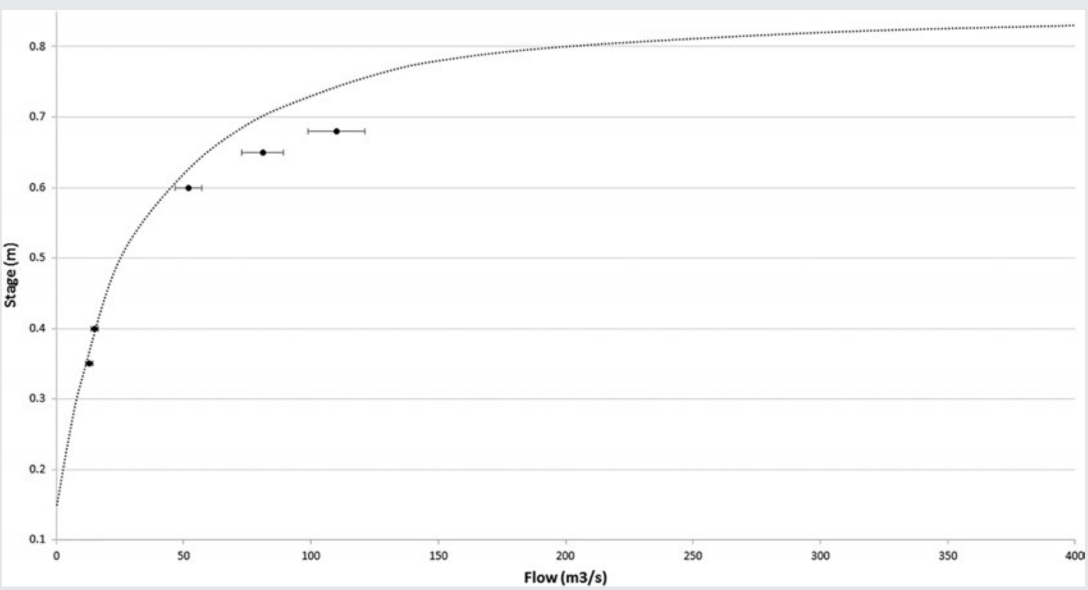


Figure 8.16 Gaugings plotted outside the current rating curve, requiring a new rating curve to be drawn. In this case at higher flows the current rating curve (dashed line) underpredicting the flow, suggesting the river-bed has eroded (bed degradation).

points at the top of the curve but will have different points at the lower end.

The points at the top of a rating curve are normally the hardest to obtain because floods occur infrequently and they are difficult to measure (both from a technical and a health and safety perspective). However, it is often those floods that are the most important to obtain measurements for as the discharge numbers become vital for flood frequency analysis. The infrequent nature of floods means that sometimes the rating curve cannot be fully drawn for several years after a new monitoring station has been established. When the top of a curve has finally been drawn it may mean altering flow records for several years previously with the new stage–discharge relationship.

The final part of applying a rating curve change is to decide when it applies from so that the flow record can be adjusted. For a flood event the rating change is normally attributed to the high point of the hydrograph (i.e. peak discharge) although in reality it is often smoothed over several hours to avoid a step change in flow. For aquatic weed growth the change may have to be smoothed over several days or weeks.

Measuring stage height

The measurement of stage has traditionally been done using a float and counterweight within a stilling well (see Figure 8.9). A float on the surface of the water within the stilling well moves up and down with the height of the river with a wire connected to the float going over a pulley at the top of the stilling well and being attached to a counterweight. Originally the pulley was connected through a small gearing mechanism to a pen marking a rotating chart so that the stage height was continuously recorded (until the chart or pen ran out!). Nowadays an electronic encoder translates the position of the pulley into a stage height that is recorded electronically in a logging device.

A stilling well with a good float and counterweight system is the gold standard for measuring stage. The stilling well is often combined with a logger and telemetry system (for sending the information from the field back to an office via radio or telephone signal) sitting on top of the well (see Figure 8.10). However, river banks can be a hostile environment and there is a financial vulnerability in having a lot of expensive equipment sitting in a position where they could be washed away during a flood. For this reason other, less accurate techniques can be used for measuring stage height. Three of the most common are: a gas bubbler, a pressure transducer, or a capacitance probe.

A gas bubbler is a small tube containing air that can be run down a river bank so that the end of the gas line is sitting on the bottom of the river. Air is forced slowly down the gas line and a measurement taken of the force required for a bubble of air to leave the end of the line. The deeper the water the harder it is to force the air bubble out of the gas line which the measuring device records. The gas bubbler and pressure sensing device can be a long way away from the river itself (e.g. 20–50 m up a bank) and well away from the destructive power of a flood.

A pressure transducer is rather like a barometer; it has a sensor in the end that sits at the bottom of a river (or lake, or groundwater well) and measures the pressure of water above it. The deeper the water the greater the pressure recorded. This pressure is translated into an electronic signal which is transmitted via a wire to the logging device which again may be well away from the river bank. A complication of using pressure transducers is that there is the pressure of water above the sensor but there is also atmospheric pressure which varies with the weather. Some pressure transducers require a coincident measurement of atmospheric pressure with an adjustment to be made later but there are now self-contained pressure transducers that automatically adjust for atmospheric pressure.

Capacitance probes measure the depth of water by measuring how easily electricity is transmitted between two sensors. The more (i.e. deeper) water present, the easier for the electrical current to be transmitted. This sounds simple enough and there are capacitance probes available that measure up to 10 m in depth. Care needs to be taken in interpreting the results from rivers as the ability to transmit electricity is also affected by the water chemistry. If this changes through a pollution incident or from differing flow sources during a flood then the capacitance probe may have a different calibration curve than for less saline water.

Flumes and weirs

Flumes and weirs utilise the stage–discharge relationship described above but go a step further towards providing a continuous record of river discharge. If we think of stream discharge as consisting of a river velocity flowing through a cross-sectional area (as in the velocity profile method) then it is possible to isolate both of these terms separately. This is what flumes and weirs, or *stream gauging structures*, attempt to do.

The first part to isolate is the stream velocity. The way to do this is to slow a stream down (or, in some rare cases, speed a stream up) so that it flows with constant velocity through a known cross-sectional area. The critical point is that in designing a flume or weir the river flows at the same velocity (or at least a known velocity) through the gauging structure irrespective of how high the river level is. Although this seems counter-intuitive (rivers normally flow faster during flood events) it is achievable if there is an area prior to the gauging structure that slows the river down: a stilling pond.

The second part of using a gauging structure is to isolate a cross-sectional area. To achieve this, a rigid structure is imposed upon the stream so that it always flows through a known cross-sectional area. In this way a simple measure of stream height through the gauging structure will give the cross-sectional area. Stream height is normally derived through a stilling well, as described in Figure 8.9, except in this case there is a regular cross-sectional area.

Once the velocity and cross-sectional area are kept fixed, the rating curve can be derived through a mixture of experiment and hydraulic theory. These relationships are normally power equations dependent on the shape of cross-sectional area used in the flume or weir. There is an international standard for manufacture and maintenance of weirs (ISO 1438) that set out theoretical ratings curves for different types of structures. The general formula for a V notch weir is shown in Equation 8.3.

$$Q = 0.53 \cdot \sqrt{2g} \cdot C \cdot \tan\left(\frac{\theta}{2}\right) \cdot h^{2.5} \qquad (8.3)$$

Where Q is discharge (m^3/s); g is the acceleration due to gravity (9.81 m/s^2); C is coefficient of discharge (see Figure 8.17); θ is the angle of V-notch (°); h is the height of water or stage (m). The coefficient of discharge can be estimated from Figure 8.17 for a certain angle of V-notch. For a 90° V-notch the coefficient of discharge is 0.578 and the rating equation becomes:

$$Q = 1.366h^{2.5} \qquad (8.4)$$

There is a similar type of equation for rectangular weirs, based on the width of the rectangular exit and another version of the coefficient of discharge relationship.

The shape of cross-sectional area is an important consideration in the design of flumes and weirs. The shape of permanent structure that the river flows through is determined by the flow regime of the river and the requirements for the streamflow data. A common shape used is based on the V-structure (see Figure 8.18). The reason for this is that when river levels are low, a small change in river flow will correspond to a significant change in stage (measured in the stilling well). This sensitivity to low flows makes data from this type of flume or weir particularly suitable for studying low flow hydrology. It is important that under high flow conditions the river does not overtop the flume or weir structure. The V shape is convenient for this also because as discharge increases

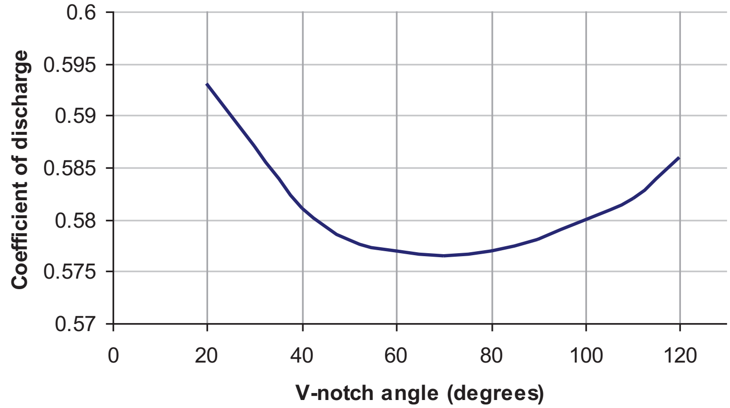


Figure 8.17 Coefficient of discharge for V-notch weirs (ISO 1438).

Figure 8.18 A V-notch weir. The water level in the pond behind the weir is recorded continuously.

the cross-sectional area flowed through increases in a non-linear fashion. The angle of the V-notch will vary depending on the size of stream being measured and the sensitivity required (90° and 120° V-notch weirs are both commonly used).

One of the difficulties in maintaining gauging structures is that by slowing the river down in a stilling pond any sediment being carried by the water may be deposited (see Hjulstrom curve in Chapter 10), which with time will fill the stilling pond and lessen its usefulness. Because of this the stilling pond needs to be dredged regularly, particularly in a high energy environment such as mountain streams. To overcome this difficulty there is a design of trapezoidal flume that speeds the stream up rather than slows it down (see Figure 8.19). The stream is forced to go down a steep section immediately prior to the gauging structure. In this way any sediment is flushed out of the weir, removing the need for regular dredging. This is really only possible for small streams as the power of large rivers at high velocities would place enormous strains on the gauging structure.

The difference between flumes and weirs

Although flumes and weirs perform the same function – measuring stream discharge in a continuous fashion – they are not the same. In a weir the water is forced to drop over a structure (the weir – Figure 8.18) in the fashion of a small waterfall. In a flume (Figure 8.19) the water passes through the structure without having a waterfall at the end.

Continuous flow gauging

The ADCP or ultrasonic flow gauging described earlier also has some capability to provide continuous flow measurement. Sideways, or upwards looking ADCP instruments mounted on a fixed structure beside a river will provide a continuous

Figure 8.19 A trapezoidal flume. The stream passes through the flume and the water level at the base of the flume is recorded continuously.

measurement of velocity passing the point. The cross-sectional area of the stream is also being collected but this is less reliable. In a fixed channel like under a bridge or an urban channel the ADCP may be able to measure depth to the bottom of the channel and the opposite side. However, for a normal river channel where the river spreads out extensively during floods it is more difficult to get an accurate measurement of cross sectional area. Probably the greatest potential for continuous ultrasonic flow gauging is in a channel affected by weed growth (see Case Study) where the velocity is being affected by a constricted channel area.

ESTIMATING STREAMFLOW

In the past 30 years, probably the greatest effort in hydrological research has gone into creating numerical models to simulate streamflow. With time these have developed into models simulating all the processes in the hydrological cycle so that far more than just streamflow can be estimated. However, it is often streamflow that is seen as the end-product of a model, a reflection of the importance streamflow has as a hydrological parameter. These models are described in Chapter 9; this section concentrates on direct estimates of streamflow.

Physical or geomorphological estimation techniques

The geomorphological approach to river systems utilises the idea that the river channel is in equilibrium with the flow regime. This suggests that measures of the channel (e.g. depth/width ratio, **wetted perimeter**, height to **bankfull discharge**) can be used to estimate the streamflows in both a historical and contemporary sense. Wharton (1995) provides a review of these different techniques. This is not a method that can be used to estimate the discharge in a river at one particular time, but it can be used to estimate parameters such as the mean annual **flood**. Important parameters to consider are the stream diameter, wetted perimeter and average depth. This is particularly for the area of channel that fills up during a small flooding event: so-called bankfull discharge.

It is possible to estimate the average velocity of a river stretch using a kinematic wave equation such as Manning's (Equation 8.5).

$$v = \frac{k \cdot R^{2/3} \cdot \sqrt{s}}{n} \tag{8.5}$$

Where v is velocity (m/s); k is a constant depending on which units of measurement are being used (1 for SI units, 1.49 for Imperial); R is the **hydraulic radius** (m); s is the slope (m/m); and n is the Manning roughness coefficient. Hydraulic radius is the cross-sectional area of a river (m^2) divided by the wetted perimeter (m). In very wide channels

Table 8.2 Chezy roughness coefficients for some typical streams

Type of channel	*Chezy roughness coefficient for a hydraulic radius of 1 m*
Artificial concrete channel	71
Excavated gravel channel	40
Clean regular natural channel <30 m wide	33
Natural channel with some weeds or stones <30 m wide	29
Natural channel with sluggish weedy pools <30 m wide	14
Mountain streams with boulders	20
Streams larger than 30 m wide	40

Source: Adapted from Richards (1982)

this can be approximated as mean depth (Goudie et al. 1994). The Manning roughness coefficient is estimated from knowledge of the channel characteristics (e.g. vegetation and bed characteristics) in a similar manner to Chezy's roughness coefficient in Table 8.2. Tables of Manning roughness coefficient can be found in Richards (1982), Chow et al. (1988), Goudie et al. (1994), and in other fluvial geomorphological texts.

Dilution gauging

Dilution gauging works on the principle that the more water there is in a river the more it will dilute a solute added into the river. There is a well-established relationship between the amount of the tracer found naturally in the stream (C_o), the concentration of tracer put into the stream (C_t), the concentration of tracer measured downstream after mixing (C_d), and the stream discharge (Q). The type of tracer used is dependent on the equipment available; the main point is that it must be detectable in solution and non-harmful to the aquatic flora and fauna. A simple tracer that is often used is a solution of table salt (NaCl), a conductivity meter being employed to detect the salt solution.

There are two different ways of carrying out dilution gauging that use slightly different equations. The first puts a known volume of tracer into the river and measures the concentration of the 'slug' of tracer as it passes by the measurement point. This is referred to as gulp dilution gauging. The equation for calculating flow by this method is shown in Equation 8.6.

$$Q = \frac{C_t V}{\sum (C_d - C_0)\Delta t} \tag{8.6}$$

Where Q is the unknown streamflow, C is the concentration of tracer either in the slug (t), downstream (d), or background in the stream (0); Δt is the time interval. The denominator of this equation is the sum of measured concentrations of tracer downstream.

The second method uses a continuous injection of tracer into the river and measures the concentration downstream. The continuous injection method is better than the slug injection method as it measures the concentration over a greater length of time however it requires a large volume of the tracer. Using the formula listed below the stream discharge can be calculated using Equation 8.7.

$$Q = q\frac{C_t - C_d}{C_d - C_0} \tag{8.7}$$

Where q is the flow rate of the injected tracer (i.e. injection rate) and all other terms are as for the gulp injection method.

Probably the most difficult part of dilution gauging is calculating the distance downstream between where the tracer is injected and the river concentration measuring point (the mixing distance). This can be estimated using Equation 8.8.

$$L = 0.13 C_Z \left(\frac{0.7 C_Z + w}{g}\right)\left(\frac{w^2}{d}\right) \tag{8.8}$$

Where L = mixing distance (m)
C_Z = Chezy's **roughness coefficient** (see Table 8.2)
w = average stream width (m)
g = gravity constant (≈9.8 m/s^2)
d = average depth of flow (m)

ESSAY QUESTIONS

1 Why might a hydrograph be best described as estimated rather than measured?

2 Describe how an acoustic Doppler current profiling (ADCP) method works and what advantages it offers over more traditional streamflow measuring techniques.

3 Using the example of a stream or river near you (or that you know well) describe the challenges involved in maintaining a rating curve at a measurement site.

4 Why might a measured streamflow at a site not fit on an established rating curve for the river?

FURTHER READING

Herschy R.W. (ed.) (1999) *Hydrometry: Principles and practices* (2nd edition). J. Wiley & Sons, Chichester.

ISO 748:2007 *Hydrometry – measurement of liquid flow in open channels using current meters or floats*. Available at <http://csmgeo.csm.jmu.edu/geollab/Whitmeyer/IrelandDocuments/Hydrometry_Measurement_2007.pdf>.

Maidment, D.R. (1993) *Handbook of hydrology*. McGraw-Hill, New York.

9

STREAMFLOW ANALYSIS AND MODELLING

LEARNING OBJECTIVES

When you have finished reading this chapter you should have:

- An understanding of what different hydrological techniques are used for.
- A knowledge of hydrograph analysis (including the unit hydrograph).
- A knowledge of how to derive and interpret flow duration curves.
- A knowledge of how to carry out frequency analysis, particularly for floods.
- An understanding of the aims of hydrological modelling and different strategies to achieve those aims.

One of the most important tasks in hydrology is to analyse streamflow data. These data are continuous records of discharge, frequently measured in permanent structures such as flumes and weirs (see Chapter 8). Analysis of these data provides us with three important features:

- description of a flow regime
- potential for comparison between rivers, and
- prediction of possible future river flows.

There are well-established techniques available to achieve these, although they are not universally applied in the same manner. This chapter sets out three important methods of analysing streamflow: hydrograph analysis, flow duration curves and frequency analysis. These three techniques are explained with reference to worked examples, all drawn from the same data set. The use of data from within the same study area is important for comparison between the techniques.

HYDROGRAPH ANALYSIS

A hydrograph is a continuous record of stream or river discharge (see Figure 7.1). It is a basic working unit for a hydrologist to understand and interpret. The shape of a hydrograph is a response from a particular catchment to a series of unique conditions, ranging from the underlying geology and catchment shape to the antecedent wetness, rainfall intensity and storm duration. The temporal and spatial variations in these underlying conditions make it highly unlikely that two hydrographs will ever be the same. Although there is great variation in the shape of a hydrograph there are common characteristics of a storm hydrograph that can be recognised. These have been described at the start of Chapter 7 where terms such as *rising limb*, *falling limb*, *recession limb* and *baseflow* are explained.

Hydrograph separation

The separation of a hydrograph into baseflow and stormflow is a common task, although never easy. The idea of **hydrograph separation** is to distinguish between stormflow and baseflow so that the amount of water resulting from a storm can be calculated. Sometimes further assumptions are made concerning where the water in each component has come from (i.e. groundwater and overland flow) but, as explained in Chapter 7, this is controversial.

The simplest form of hydrograph separation is to draw a straight, level line from the point where the hydrograph starts rising until the stream discharge reaches the same level again (dashed line in Figure 9.1). However, this is frequently problematic as the stream may not return to its pre-storm level before another storm arrives. Equally, the storm may recharge the baseflow enough so that the level is raised after the storm (as shown in Figure 9.1).

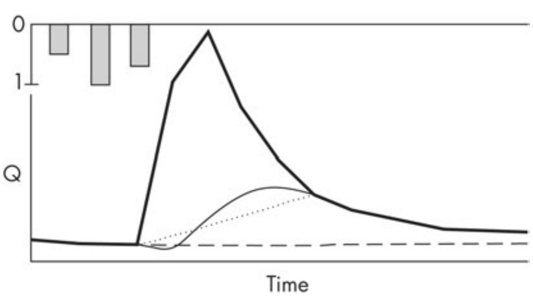


Figure 9.1 Hydrograph separation techniques. See text for explanation.

To overcome the problem of a level baseflow separation, a point has to be chosen on the receding limb where it is decided that the discharge has returned to baseflow. Exactly where this point will be is not easy to determine. By convention, the point is taken where the recession limb fits an exponential curve. This can be detected by plotting the natural log (ln) of discharge (Q) and noting where this line becomes straight, but even using this technique requires an element of subjective judgement. The line drawn between the start and 'end' of a storm may be straight (dotted line, see Figure 9.1) or curved (thin solid line, see Figure 9.1) depending on the preference of the hydrologist – Arnold et al. (1995) provides a summary of different automated techniques.

In very large catchments, Equation 9.1 has been applied to derive the time where stormflow ends (Linsley et al. 1975). This is the fixed time method which gives the time from peak flow to the end of stormflow (τ):

$$\tau = 0.8278A^2 \qquad (9.1)$$

Where A is the drainage area (km^2) and τ is in days.

The problem with hydrograph separation is that the technique is highly subjective. There is no physical reasoning why the 'end' of a storm should be when the recession limb fits an exponential curve; it is pure convention. Equally the shape of the curve between start and 'end' has no physical reasoning. It does not address the debate covered in Chapter 7: where does the stormflow water come from? Furey and Gupta (2001) provide a hydrograph separation technique that ties into physical characteristics of a catchment and therefore is not as subjective as other techniques, although it still requires considerable interpretation by the user. The general view is that whatever method is used, it should be applied consistently. What hydrograph separation does offer is a means of separating stormflow from baseflow, something that is needed for the use of the unit hydrograph (see pp. 178–181), and may be useful for hydrological interpretation and description.

The unit hydrograph

The idea of a **unit hydrograph** was first put forward by Leroy Sherman, an American engineer working in the 1920s and 1930s. The idea behind the unit hydrograph is simple enough, although it is a somewhat tedious exercise to derive one for a catchment. The fundamental concept of the unit hydrograph is that the shape of a storm hydrograph is determined by the physical characteristics of the catchment. The majority of those physical characteristics are static in time, therefore if you can find an average hydrograph for a particular storm size then you can use that to predict other storm events. In short: two identical rainfall events that fall on a catchment with exactly the same antecedent conditions should produce identical hydrographs.

With the unit hydrograph, a hydrologist is trying to predict a future storm hydrograph that will result from a particular storm. This is particularly useful as it gives more than just the peak runoff volume and includes the temporal variation in discharge.

Sherman (1932) defines a unit hydrograph as ‘the hydrograph of surface runoff resulting from **effective rainfall** falling in a unit of time such as 1 hour or 1 day’. The term effective rainfall is taken to be that rainfall that contributes to the storm hydrograph – it is what remains after all ‘losses’ such as interception, depression storage, soil surface wetting etc. It is the proportion of the rainfall that actually drives the stormflow response (Figure 9.2). This is often assumed to be the rainfall that does not infiltrate the soil and moves into the stream as overland flow. This is infiltration excess or Hortonian overland flow. Sherman’s ideas fitted well with those of Horton. Sherman assumed that the ‘surface runoff is produced uniformly in space and time over the total catchment area’.

All these calculations can be undertaken easily using a spreadsheet package such as Microsoft Excel. For the purposes of explanation we will assume familiarity with Excel, but it should be relatively easy to establish equivalent functions in other spreadsheet packages. For all the techniques in this chapter, we recommend that you download the relevant data and

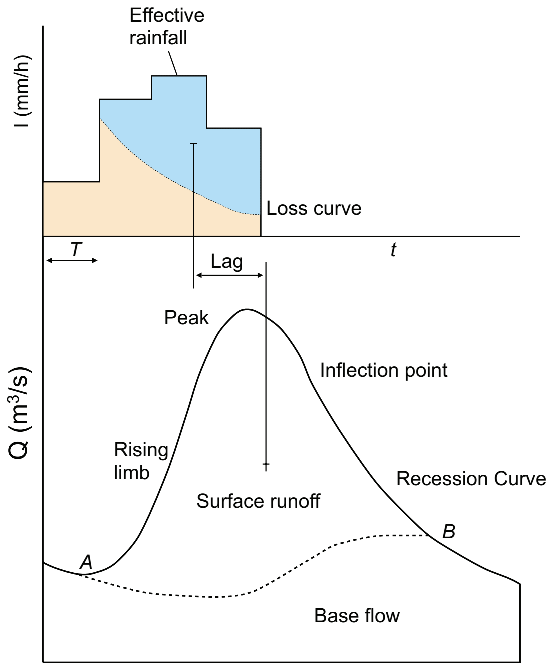


Figure 9.2 The concept of effective rainfall and its relationship to the stormflow hydrograph.

work through the steps – in this way you will build a practical understanding of the techniques. For each technique in this chapter there is an explanation of the steps required and also a worked example based on the downloadable data.

Deriving the unit hydrograph

Step 1: Storm selection

Take historical rainfall and streamflow records for a catchment and separate out a selection of typical single-peaked storm hydrographs (Figure 9.3). In this example, we will consider one storm, but conventionally you would identify five to seven storms in order to be sure you were getting an ‘average’ hydrograph response. It is important that they are separate storms as the method assumes that one runoff event does not affect another. Once you have identified your candidate storms, copy and

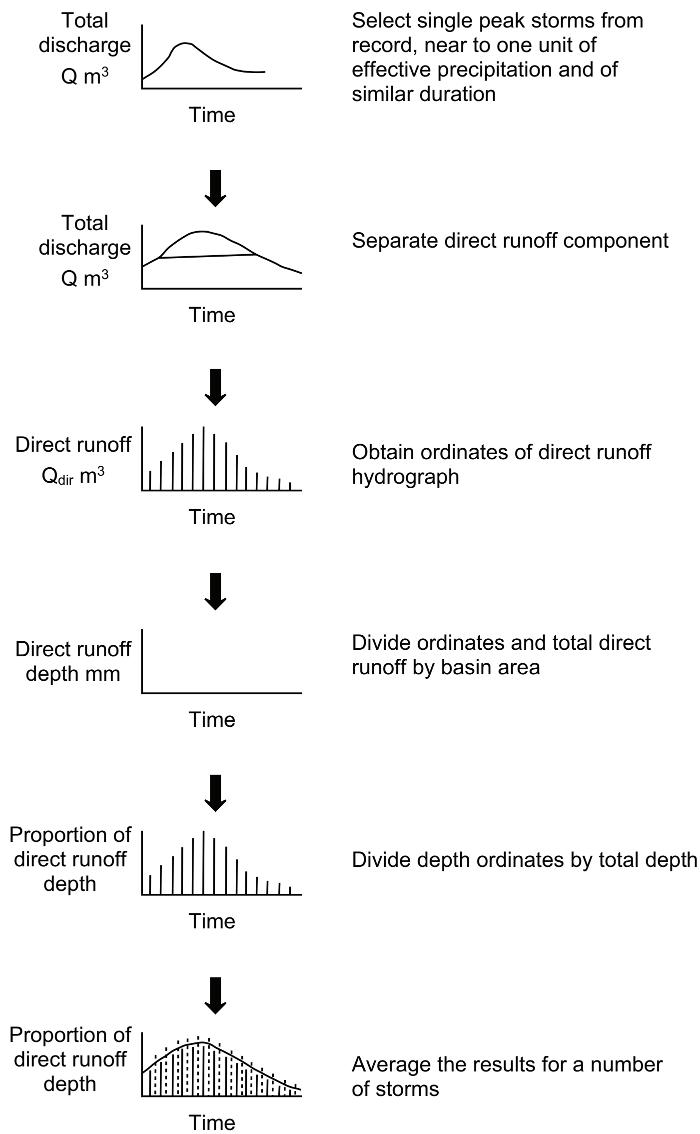


Figure 9.3 Steps in deriving a unit hydrograph for a catchment.

paste the rainfall and flow data for one storm into a new worksheet. From now on, you will be working with this storm and hydrograph independently but will go back and repeat the steps below for each one of the five to seven candidate storms you have identified.

Step 2: Hydrograph (baseflow) separation

The next step is to establish what the stormflow component of the hydrograph is. To do this you need to separate the baseflow from the stormflow; that is, hydrograph separation (see p. 177). You will need to decide which method to use and later apply this approach consistently for all storms. Once you have decided the 'start' and 'end' points of the baseflow separation line, you will need to use linear interpolation (e.g. the FORECAST function in Excel) to calculate the baseflow at each time step along the baseflow separation line. As you already have the total recorded flow in the hydrograph, subtracting the interpolated baseflow value at each time step will leave you with the stormflow component of the hydrograph for each time step – in other words what share of the total discharge at each time step was stormflow. At this point you are still working in the original units of the hydrograph (m^3/s).

Step 3: Calculate the total stormflow volume

In the next step you need to work out the total volume of water that contributed to the storm. Referring back to Figure 7.3, this is of course the area between the baseflow separation line and the hydrograph. This can be done either by measuring the area under the stormflow hydrograph or as an integral of the curve. In Excel you could take the average stormflow (m^3/s) for each time step and multiply it by the time step duration (s). This would give you the stormflow volume (m^3) in each time step, and by summing these across the storm duration you would have an estimate of the total storm flow volume.

Step 4: Calculate the water equivalent depth

In the unit hydrograph method, we assume that the runoff is generated evenly across the catchment. So what rainfall would be required to generate the surface runoff volume we have calculated? If you then divide the total volume in the storm by the catchment area, you will have expressed the runoff as a **water equivalent depth**. In doing so you will need to remember to check your unit conversions; because it relates to the driving rainfall, water equivalent depth is expressed in mm. If this is assumed to have occurred uniformly over space and time within the catchment then you can assume it is equal to the effective rainfall. This is an important assumption of the method: that the effective rainfall is equal to the water equivalent depth of storm runoff. It is also assumed that the effective rainfall all occurred during the height of the storm (i.e. during the period of highest rainfall intensity). That period of high rainfall intensity becomes the time for the unit hydrograph.

Step 5: Calculate the ordinates of the unit hydrograph

The unit hydrograph is defined as the stormflow that results from one unit of effective rainfall. In Step 2 you calculated the stormflow discharge (m^3/s) at each time step by subtracting the baseflow from the total discharge. In Step 4 you calculated the effective rainfall. By 'ordinates' of the unit hydrograph we mean the values of the unit hydrograph at each time step. To derive this you need to divide the values of stormflow (m^3/s) (i.e. each value on the storm hydrograph – Step 2) by the amount of effective rainfall (mm) (from Step 4) to give the unit hydrograph. This gives the unusual units of the y-axis of the unit hydrograph ($m^3/s/mm$) – but think about what this means – this is the discharge (m^3/s) *per* millimetre of effective rainfall during the time unit. The resultant unit hydrograph looks a bit like the original hydrograph but has time since the storm as the x-axis and storm discharge per mm of effective rainfall ($m^3/s/mm$) as the y-axis.

Step 6: Creating an average hydrograph

Steps 1 to 5 above related to one storm only – but we wouldn't know whether this would have been a representative storm or not. To obtain a more representative hydrograph, Steps 2 to 5 would need to be repeated for all candidate hydrographs selected in Step 1 and an average unit hydrograph obtained by merging the curves together. This is achieved by averaging the value of stormflow for each unit of time for every derived unit hydrograph. The result would be a more robust estimate of the stormflow that is generated for each mm of effective rainfall that occurs. This could be thought of as the 'stormflow signature' of the catchment which can then be used to estimate the stormflow hydrograph arising from a storm of any other depth or duration – this is the value of the method (see next section for how this is done). It is also possible to derive different unit hydrographs for different rain durations and intensities, but this is not covered here (see Maidment (1992), or Shaw et al. (2011) for more details).

Using the unit hydrograph

The unit hydrograph obtained from the steps described here theoretically gives you the runoff that can be expected per mm of effective rainfall in 1 hour. In order to use the unit hydrograph for predicting a storm it is necessary to estimate the 'effective rainfall' that will result from the storm rainfall. This is not an easy task and is one of the main hurdles in using the method. In deriving the unit hydrograph the assumption has been made that 'effective rainfall' is the rainfall which does not infiltrate but is routed to the stream as overland flow (Hortonian). The same assumption has to be made when utilising the unit hydrograph. To do this it is necessary to have some indication of the infiltration characteristics for the catchment concerned, and also of the antecedent soil moisture conditions. The former can be achieved through field experimentation and the latter through the use of an antecedent precipitation index (API). Engineering textbooks give examples of how to use the API to derive effective rainfall. The idea is that antecedent soil moisture is controlled by how long ago rain has fallen and how large that event was. The wetter a catchment is prior to a storm, the more effective rainfall will be produced.

Once the effective rainfall has been established, it is a relatively simple task to add the resultant unit hydrographs together to form the resultant storm hydrograph. Before doing this though, it is important to understand the three principles upon which the method is based (Figure 9.4):

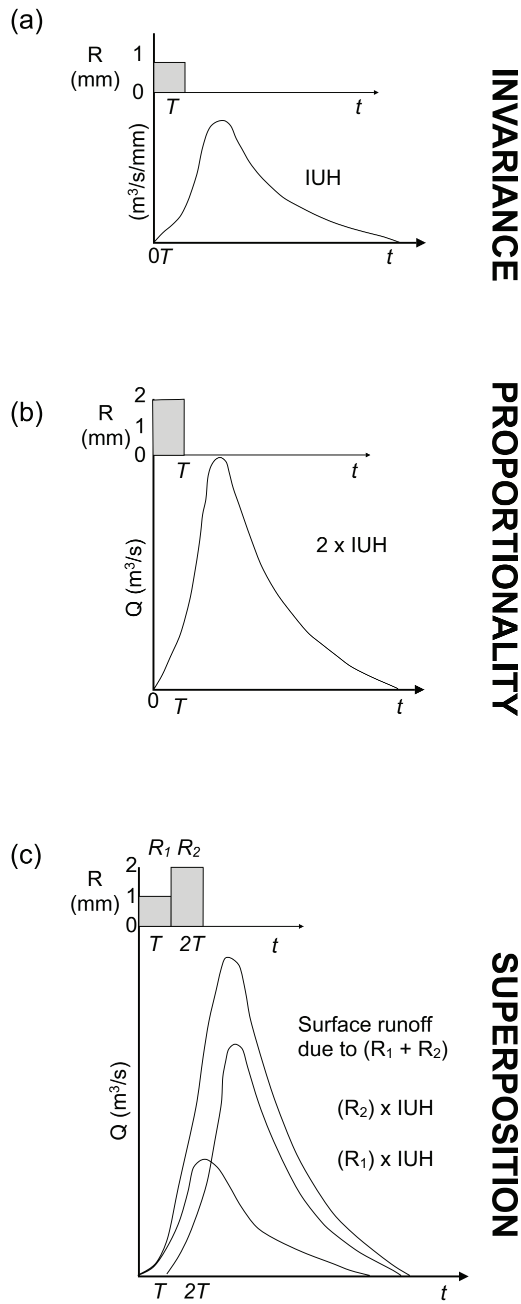


Figure 9.4 Illustration of the principles underpinning application of the unit hydrograph. Adapted from Shaw et al. (2011).

The principle of invariance

Once you have derived your unit hydrograph ordinates, you have established a relationship between effective rainfall of 1 mm and how this generates runoff over time. This principle assumes that this relationship holds true all the time and does not vary by storm or seasonally (i.e. it is invariant).

The principle of proportionality

This principle assumes a direct relationship between the magnitude of effective rainfall and the runoff generated. In other words, if you double the effective rainfall, you can simply double the magnitude of your unit hydrograph ordinates; doubling the effective rainfall will double the stormflow.

The principle of superposition

Whilst the unit hydrograph is the amount of stormflow generated for 1 mm falling in 1 hour, this principle allows us to determine what happens when we have storms that last for more than 1 hour. This is implemented by drawing the hydrograph that results from the first hour and then drawing the second hydrograph that would result from whatever effective rainfall occurred in the second hour – but importantly – offset by 1 hour, as the response only started in the second hour. If the storm was 3 hours long you would have three hydrographs, each offset by 1 hour with respect to each other. The total hydrograph is estimated by simply adding all three together at each time step, a process known as **convolution** of the unit hydrograph. This is best understood by a practical example – see the worked example as an illustration of how this is done.

Limitations of the unit hydrograph

As indicated above, the unit hydrograph has several assumptions that at first appearance would seem to make it inapplicable in many situations. The assumptions can be summarised as:

- The runoff that makes up stormflow is derived from infiltration excess (Hortonian) overland flow. As described in Chapter 7, this is not a reasonable assumption to make in many areas of the world.
- The surface runoff occurs uniformly over the catchment because the rainfall is uniform over the catchment. Another assumption that is difficult to justify.
- The relationship between effective rainfall and surface runoff does not vary with time (i.e. the hydrograph shape remains the same between the data period of derivation and prediction). This would assume no land-use change within the catchment, as this could well affect the storm hydrograph shape.

Given the assumptions listed above it would seem extremely foolhardy to use the unit hydrograph as a predictive tool. However, the unit hydrograph has been used successfully for many years in numerous different hydrological situations. In fact, many approaches to the estimation of hydrographs across the world are based on this fundamental thinking. It is a very simple method of deriving a storm hydrograph from a relatively small amount of data. The fact that it does work (i.e. produces meaningful predictions of storm hydrographs), despite being theoretically flawed, would seem to raise questions about our understanding of hydrological processes. The answer to why it works may well lie in the way that it is applied, especially the use of effective rainfall. This is a nebulous concept that is difficult to describe from field measurements. It is possible that in moving from actual to effective rainfall, there is a blurring of processes that discounts some of the assumptions listed above. The unit hydrograph is a black box model of stormflow (see end of this chapter) and as such hides many different processes within. The simple concept that the hydrograph shape is a reflection of the static characteristics and all the dynamic processes going on in a catchment makes it highly applicable but less able to be explained in terms of hydrological theory.

The synthetic unit hydrograph

The **synthetic unit hydrograph** is an attempt to derive the unit hydrograph from measurable catchment characteristics rather than from flow data. This is highly desirable as it would give the opportunity to predict stormflows when having no historical streamflow data; a common predicament around the world. The Institute of Hydrology in the UK (now the Centre for Ecology & Hydrology – CEH) carried out an extensive study into producing synthetic unit hydrographs for catchments, based on factors such as the catchment size, degree of urbanisation, etc. (NERC 1975). They produced a series of multiple regression equations to predict peak runoff amount, time to peak flow, and the time to the end of the recession limb based on the measurable characteristics. Although this has been carried out relatively successfully it is only applicable to the UK as this is where the derivative data was from. In another climatic area the hydrological response is likely to be different for a similar catchment. The UK is a relatively homogeneous climatic area with a dense network of river flow gauging, which allowed the study to be carried out. In areas of the world with great heterogeneity in climate and sparse river monitoring it would be extremely difficult to use this approach.

Worked example of the unit hydrograph

The Tanllwyth is a small (0.98 km^2) headwater tributary of the River Severn in mid-Wales. The catchment is monitored by the Centre for Ecology and Hydrology (formerly the Institute of Hydrology) as part of the Plynlimon catchment experiment. For this example, a storm was chosen from July 1982 as it is a simple single-peaked hydrograph (see Figure 9.3).

Baseflow separation was carried out by using a straight-line method. The right-hand end of the straight line (shown as a broken line in Figure 9.6) is where the receding limb of the hydrograph became exponential.

All of the flow above the broken line in Figure 9.6 was then divided by the effective rainfall to derive the unit hydrograph in Figure 9.7 below.

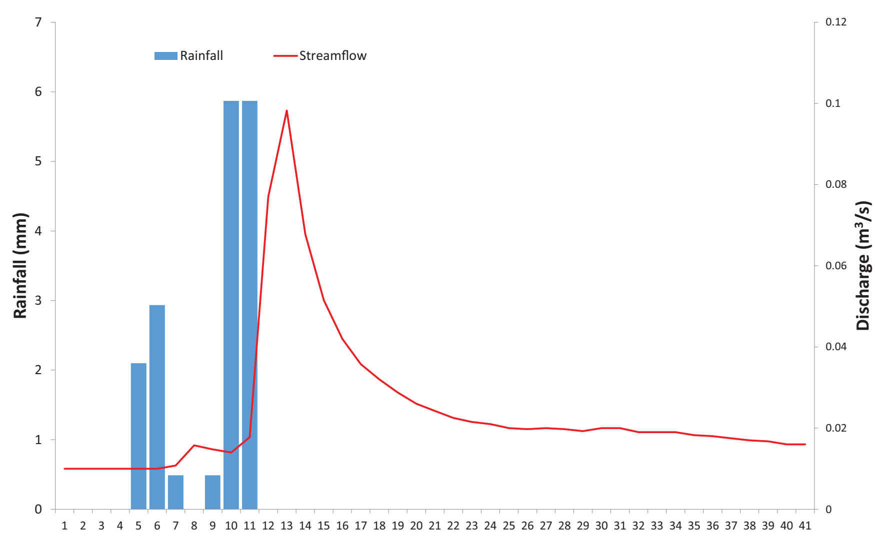


Figure 9.5 A simple storm hydrograph (July 1982) from the Tanllwyth catchment.

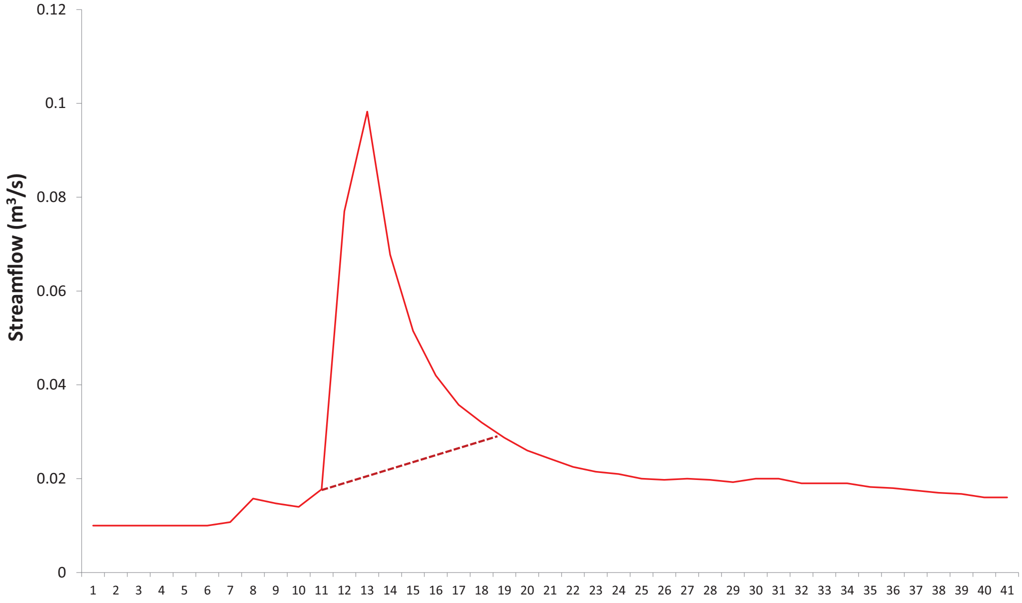


Figure 9.6 Baseflow separation.

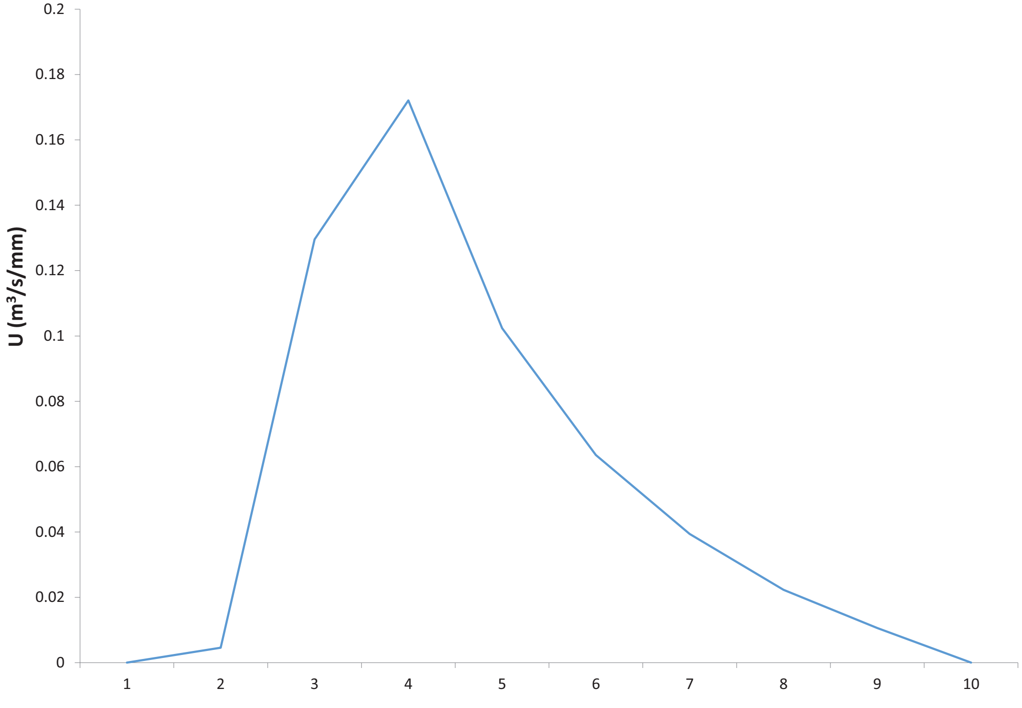


Figure 9.7 The unit hydrograph for the Tanllwyth catchment.

To apply the unit hydrograph to a small storm, hydrographs were added together for each amount of effective rainfall. The resultant total hydrograph is shown as the purple line in Figure 9.8. The discharge values in the simulated hydrograph are much larger than those in the original storm hydrograph despite what appears to be a smaller storm. This is because the simulated hydrograph is working on effective rainfall rather than actual rainfall. Effective rainfall is the rain that doesn't infiltrate and is theoretically available for storm runoff. A low effective rainfall value may represent a high actual rainfall value.

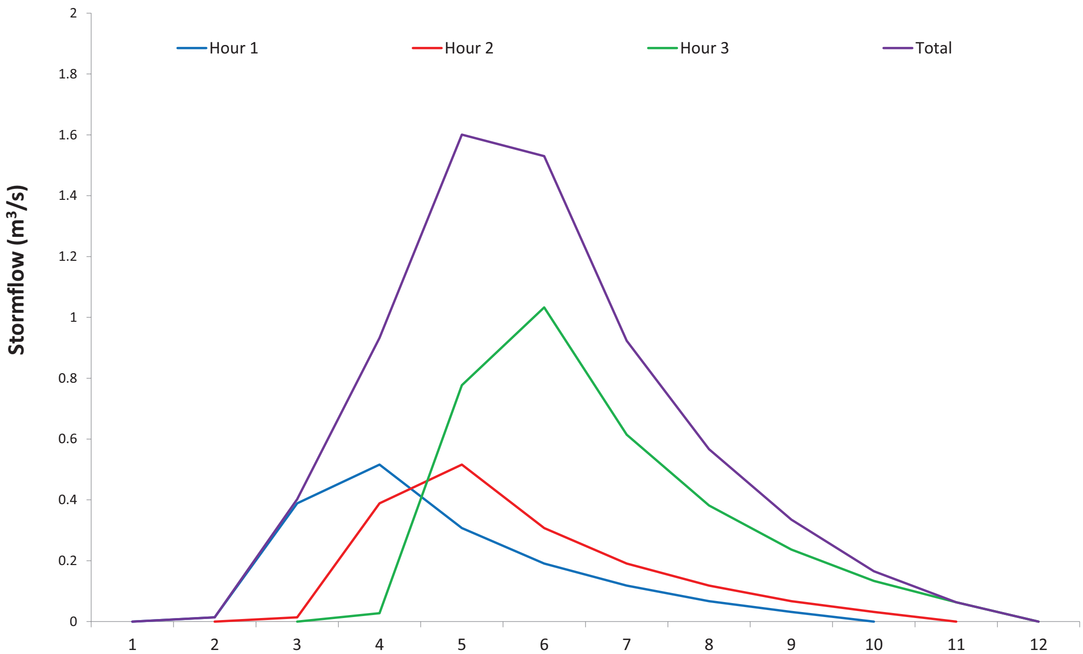


Figure 9.8 Applying the unit hydrograph to a small storm (3 mm in the first hour, 6 mm in the second hour and 6 mm in the third hour). The different lines represent the flow from each of the hourly rainfalls (blue first, then red, then green). The purple line is the total discharge i.e. the sum of the three lines.

FLOW DURATION CURVES

An understanding of how much water is flowing down a river is fundamental to hydrology. Of particular interest for both flood and low flow hydrology is the question of how representative or 'typical' a certain flow is. This can be addressed by looking at the frequency of daily flows and some statistics that can be derived from the frequency analysis. For example, it might be of interest to know which flows occur less than 5 per cent of the time. One convenient way of showing this is a **flow duration curve**, one of the essential basic tools of a hydrologist.

Whereas a hydrograph shows the sequence of flows, in a flow duration curve the time sequence of the flows is lost, but other valuable information based on the frequency analysis is gained. Flow duration curves show on the y-axis the flow and on the x-axis the percentage of time that the flow is equalled or exceeded. The flow associated with a particular frequency is known as a 'Q-value'. The 50 per cent value or 'Q_{50}' therefore represents the median flow, an important indicator. This means that half the time flow in the river is higher than the Q_{50} discharge, and half the time it will be lower. In Figure 9.9, the value marked by a dot is the Q_{30} value – read as '30 per cent of the time the flow equals or exceeds about 0.67 m^3/s'. Other indicator statistics commonly used include the Q_{95} and Q_{99} values (indicators of low flows), whereas the Q_{10} and Q_5 are high flow indicators (flows only exceed these values 10 per cent and 5 per cent of the time, respectively).

You will also notice from Figure 9.9 that the flow duration curve appears exponential – this makes it difficult to read off the very low and very high frequency Q-values. One way of getting around this is to undertake a log transformation of the data. In our example we take the natural log (LN function in Excel) of the flows and plot these. This has the effect of straightening out the flow duration curve so it is easier to read. Values from the x-axis can be converted back to normal flow by taking the inverse of the transformed flow.

The data most commonly used to derive a flow duration curve are daily mean flows: the average flow for each day (note well that this is not the same as a mean daily flow, which is the average of a series of daily flows). To derive a representative flow duration curve the daily mean flow data are required for a long period of time, in excess of 5 years. That is not to say that a flow duration curve cannot be derived for a particular year – as in Figure 9.9 – it is

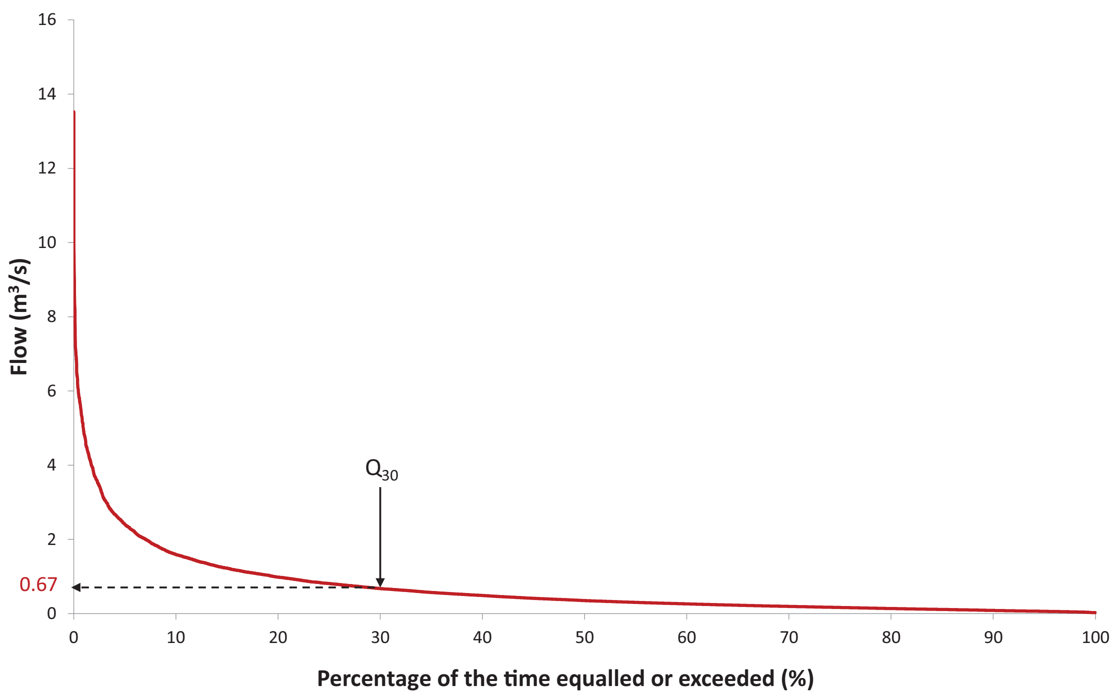


Figure 9.9 Flow duration curve for the Wye Flume (1970/71 to 1994/95). The arrow marks the Q_{30} value, the flow that is equaled or exceeded 30% of the time (0.67 m^3/s).

Source: Data from the UK National River Flow Archive (https://nrfa.ceh.ac.uk) just if you want it to be representative of the range of natural variation, you need as long a record as possible.

There are a number of methods for creating a flow duration curve, but underpinning all is essentially a frequency analysis of flows. Prior to the advent of computers this was a painstaking process. The worked example provided here shows how the flow duration curve would have been constructed historically. The example uses 26 years of data for the upper reaches of the river Wye in mid-Wales, UK (see pp. 183–185). We present this method because it is helpful in understanding the fundamentals of the process, and recommend that you do it once to help cement this understanding. Having understood the fundamental principle, we will then show you a quicker method that is available through spreadsheet functions. In practice, hydrologists normally use specialist software to calculate and display flow duration curves.

Flow duration curve

Step 1

A table is derived that has, as a first column, the frequency of occurrence of the flows for each of the flow classes that you select to represent the range of flows. This is simply the number of times across the full record the daily flow fell into that class (or range of flows). In the worked example in Table 9.1 for example, there were 250 instances of flow being less than 0.05 m^3/s. In deciding on the classes, it is important that a small class interval (or 'bin' as they are sometimes referred to in frequency analysis software packages) is used; too large an interval and information will be lost from the flow duration curve. The method for choosing the best class interval is essentially through trial and error. As a general rule, you should aim not to have more than around 10 per cent of your values within a single class interval. If you have more than this you start to lose precision in plotting. As shown in the worked example, it is not essential that the same interval is used throughout. As mentioned earlier, historically this would have been a very laborious process. With Excel it can be done relatively easily using the FREQUENCY function. The sum of all the frequency counts (9437 in the worked example) should equal the total number of observations in the record. The next column – relative frequency – is simply the respective frequency divided by the total number of observations and expressed as a percentage. So in the worked example, 2.65 per cent of flows were in the first class. The cumulative frequency is then calculated by starting from the highest flow class and then adding each successive relative frequency value – so that by the time you reach the class representing the lowest flow category you should total 100 per cent. The percentage cumulative frequency is assumed to equal the percentage of time that the flow is exceeded.

Step 2

The actual flow duration curve is created by plotting the percentage cumulative frequency on the x-axis against the mid-point of the class interval on the y-axis. Where two flow duration curves are presented on the same axes they need to be standardised for direct comparison. One way of doing this is to divide the values on the y-axis (mid-point of class interval) by the average daily flow for the record length. This makes the y-axis a percentage of the average flow (see Figure 9.10).

As mentioned earlier, the presentation of a flow duration curve may be improved by either plotting on a special type of graph paper or transforming the data. The type of graph paper often used has the x-axis transformed in the form of a known distribution such as the Gumbel or Log Pearson. A natural log transformation of the flow values (y-axis) achieves a similar effect, although this is not necessarily standard practice.

A quicker method for deriving a flow duration curve

Historically the process of constructing the flow duration curve was time consuming, and even with the benefit of the FREQUENCY function in Excel it is still laborious. A much quicker alternative is to use the PERCENTILE.EXC function in Excel. Let's say you were interested in the Q_{10} – remember this is the flow that is equalled or exceeded 10 per cent of the time, or in other words, 90 per cent of the flows are less than this. By selecting your whole data record and providing the parameter '0.9' to represent the 90th percentile, the PERCENTILE.EXC function will return the Q_{10} value. The flow duration curve can easily be created by calculating all the Q values (for the parameters 0 to 1) representing all percentiles. The flow duration curve is then just the Q value plotted against the corresponding flow. This method is of course much easier, but what is being done is hidden behind the concept of a percentile. This is the reason for illustrating the historical method first – it allows a more intuitive understanding of what lies behind a flow duration curve.

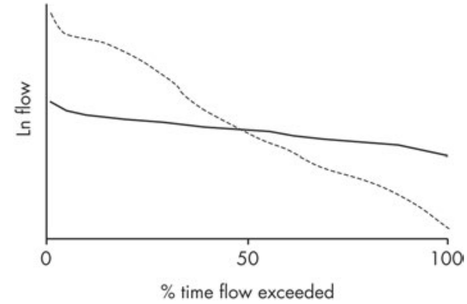


Figure 9.10 Two contrasting flow duration curves. The dotted line has a high variability in flow (similar to a small upland catchment) compared to the solid line (similar to a catchment with a high baseflow).

Interpreting a flow duration curve

The shape of a flow duration curve can tell a lot about the hydrological regime of a catchment. In Figure 9.10 two flow duration curves of contrasting shape are shown. With the dotted line there is a large difference between the highest and lowest flow values, whereas for the solid line there is far less variation. This tells us that the catchment shown by the solid line never has particularly low flows or particularly high flows. This type of hydrological response is found in limestone or chalk catchments where there is a high baseflow in the summer (groundwater derived) and high infiltration rates during storm events. In contrast the catchment shown with the dotted line has far more variation. During dry periods it has a very low flow, but responds to rainfall events with a high flow. This is characteristic of impermeable upland catchments or streamflow in dryland areas. Baseflow dominated catchments typically have a much flatter flow duration compared with flashier stormwater dominated catchments – so the slope of the flow duration curve tells us something about storage

Worked example of flow duration curve

The Wye river has its headwaters in central Wales and flows into the Severn at the head of the Severn Estuary. In its upper reaches it is part of the Plynlimon hydrological experiment run by the Institute of Hydrology (now the Centre for Ecology and Hydrology) from the early 1970s. At Plynlimon the Wye is a small (10.5 km^2) grassland catchment with an underlying geology of relatively impermeable Ordovician shale. The data used to derive a flow duration curve here are from the Upper Wye for a period from 1970 until 1995, consisting of 9,437 values of daily mean flow in cumecs (Figure 9.11).

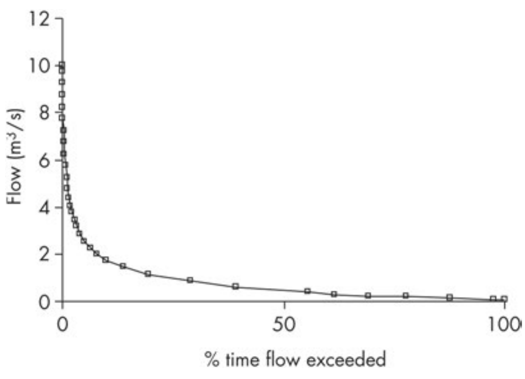


Figure 9.11 Flow duration curve for the river Wye (1970–1995 data).

The frequency analysis for the Wye gives Table 9.1.

Table 9.1 Values from the frequency analysis of daily mean flow on the upper Wye catchment.

Daily mean flow (m^3/s)	*Frequency*	*Relative frequency (%)*	*Cumulative frequency (%)*
0–0.05	250	2.65	100.00
0.05–0.1	923	9.78	97.35
0.1–0.15	927	9.82	87.57
0.15–0.2	814	8.63	77.75
0.2–0.25	708	7.50	69.12
0.25–0.3	589	6.24	61.62
0.3–0.4	881	9.34	55.38
0.4–0.5	641	6.79	46.04
0.5–0.7	958	10.15	39.25
0.7–1.0	896	9.49	29.10
1.0–1.3	553	5.86	19.60
1.3–1.6	357	3.78	13.74
1.6–1.9	222	2.35	9.96
1.9–2.1	117	1.24	7.61
2.1–2.4	127	1.35	6.37
2.4–2.7	103	1.09	5.02
2.7–3.0	71	0.75	3.93
3.0–3.3	47	0.50	3.18
3.3–3.6	42	0.45	2.68
3.6–3.9	34	0.36	2.24
3.9–4.2	33	0.35	1.88
4.2–4.5	28	0.30	1.53
4.5–5.0	28	0.30	1.23
5.0–5.5	23	0.24	0.93
5.5–6.0	23	0.24	0.69
6.0–6.5	14	0.15	0.45
6.5–7.0	7	0.07	0.30
7.0–7.5	7	0.07	0.22
7.5–8.0	3	0.03	0.15
8.0–8.5	4	0.04	0.12
8.5–9.0	2	0.02	0.07
9.0–9.5	1	0.01	0.05
9.5–10.0	3	0.03	0.04
>10.0	1	0.01	0.01
Total	9,437	100	

Note: These values form the basis of the flow duration curve in Figure 9.10

The flow duration curve is derived by plotting the percentage cumulative frequency (x-axis) against the mid-point of the daily mean flow class intervals (y-axis), except for the highest flow class where you should use the maximum flow. When this is plotted it forms the exponential shape that is normal for this type of catchment (see Figure 9.11). In order to see more detail on the curve the flow values can be logged (natural log). This is shown in Figure 9.12. The flow statistics Q_{95}, Q_{50} and Q_{10} can either be read from the graph (see Figure 9.12) or interpolated from the original frequency table (remembering to use the mid-points of the class interval). A summary of the flow statistics for the upper Wye are shown in Table 9.2.

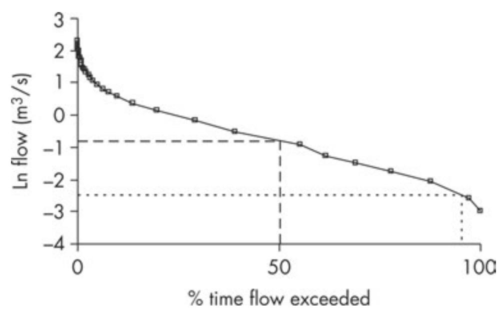


Figure 9.12 Flow duration curve for the river Wye (1970–1995 data) with the flow data shown on a natural log scale. Q_{95} (short dashes) and Q_{50} (long dashes) are shown on the flow duration curve.

Table 9.2 Summary flow statistics derived from the flow duration curve for the Wye catchment

Flow statistic	*Ln flow (m^3/s)*	*Flow (m^3/s)*
Q_{95}	– 2.48	0.084
Q_{50}	– 0.91	0.405
Q_{10}	0.56	1.75

FREQUENCY ANALYSIS

The analysis of how often an event is likely to occur is an important concept in hydrology. It is the application of statistical theory into an area that affects many people's lives, whether it be through flooding or low flows and drought. Both of these are considered here, although because they use similar techniques the main emphasis is on **flood frequency analysis**. The technique is a statistical examination of the frequency–magnitude relationship discussed in Chapters 1 and 7. It is an attempt to place a probability on the likelihood of a certain event occurring. Predominantly it is concerned with the low-frequency, high-magnitude events (e.g. a large flood or a very low river flow).

It is important to differentiate between the uses of flow duration curves and frequency analysis. Flow duration curves tell us the percentage of time that a flow is above or below a certain level. This is average data and describes the overall flow regime. Flood frequency analysis is concerned only with peak flows: the probability of a certain flood recurring. Conversely, **low flow frequency analysis** is concerned purely with the lowest flows and the probability of them recurring.

Flood frequency analysis

Flood frequency analysis is probably the most important hydrological technique and seeks to establish the probability of floods of certain magnitudes occurring, based on what has happened in the past and for which we have observed data. This brings into focus one of the major challenges in flood frequency analysis – often we don't have sufficient observed data. A second major challenge is that even if we do have reasonable data, is the past a good indication of what might happen in the future? Changes in catchments (urbanisation, impoundments) and climate change may mean that the past hydrological record is a poor indicator of future conditions – particularly when we are interested in design estimates for structures that we want to last well into the future.

Most people have heard the expression '100-year return period', but unfortunately this is widely misunderstood. For example, people who may have just endured the 100-year event may mistakenly believe that this flood would not happen for another 100 years. This is not what is meant by return period. It also does not mean that a 100-year flood must happen every 100 years. The technical definition of the return period is the long term average interval between the exceedences of a flood of a certain magnitude – or to express it a little more simply – the long term average time between two floods of a given size. The implication is that it is possible to experience two 100-year events within a few years. Consider someone living next to a river; it can be shown mathematically that that person (assuming a life span of 70 years) has a 51 per cent chance of experiencing the 100-year flood in their lifetime, and a 15 per cent chance of experiencing it twice. Misunderstanding of the concept of return period and communication of flood risk has led to the suggestion that the term 'recurrence interval' should rather be used (ICE 2001). In technical contexts we prefer to use the term **annual probability of flooding** or **APF** – we will define this more precisely later.

Flood frequency analysis is concerned with peak flows, and normally these are recorded separately to daily mean flows. In the United Kingdom, for example, gauging stations record 15-minute samples. These values are averaged over a 24-hour period to give the gauged daily flow or the daily mean flow. However, the highest flow out of the 15-minute samples is also recorded – this is the peak flow for the day. It is the daily *mean* flows that are used for the flow duration curve, but daily *peak* flows that are used for flood frequency analysis. In situations where peak flows are not recorded one can use the daily mean, provided one accepts that this would represent an underestimate of peaks. There are two different ways that a peak flow can be defined:

- the single maximum peak within a year of record giving an **annual maximum series** (sometimes abbreviated to AMAX); or

- any flow above a certain threshold value, giving a **partial duration series**, also known as **peaks over threshold** (sometimes abbreviated to POT).

Figure 9.13 shows the difference between these two data series. There are arguments for and against the use of either data series in flood frequency analysis. Annual maximum may miss a large storm event where it occurs more than once during a year (as in the 1981 case in Figure 9.13), but it does provide a continuous series of data that are relatively easy to process. The setting of a threshold storm (the horizontal line in Figure 9.13) is critical in analysis of the partial duration series, something that requires considerable experience to get right (typically one would aim for between three and five events per year). The most common analysis is on annual maximum series, the simplest form, which is described here. If the data series is longer than 10 years then the annual maxima can be used; for very short periods of record the partial duration series can be used.

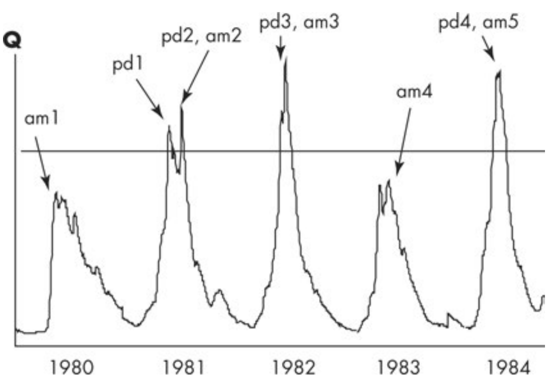


Figure 9.13 Daily flow record for the Adams river (British Columbia, Canada) during 5 years in the 1980s. Annual maximum series are denoted by 'am', partial duration series above the threshold line by 'pd'. NB In this record there are five annual maximum data points and only four partial duration points, including two from within 1981.

Source: Data courtesy of Environment Canada

The first step in carrying out flood frequency analysis is to obtain the data series (in this case annual maxima). The annual maximum series should be for as long as the data record allows. The greater the length of record the more certainty can be attached to the prediction of average recurrence interval. Many hydrological database software packages (e.g. HYDSYS) will give annual maxima data automatically, but some forethought is required on what annual period is to be used. There is an assumption made in flood frequency analysis that the peak flows are independent of each other (i.e. they are not part of the same storm). If a calendar year is chosen for a humid temperate environment in the northern hemisphere, or a tropical region, it is possible that the maximum river flow will occur in the transition between years (i.e. December/January). It is possible for a storm to last over the 31 December/1 January period and the same storm to be the maximum flow value for both years. If the flow regime is dominated by snow melt then it is important to avoid splitting the hydrological year at times of high melt (e.g. spring and early summer). To avoid this it is necessary to choose your hydrological year as changing during the period of lowest flow. This may take some initial investigation of the data. For example, because high flows are normally experienced in the northern hemisphere winter, in the UK the water or hydrological year runs from 1 October to 31 September. This means the whole wet season is contained within one water year.

All flood frequency analysis is concerned with the analysis of frequency histograms and probability distributions. Consequently the first data analysis step should be to draw a frequency histogram. It is often useful to convert the frequency into a relative frequency (divide the number of readings in each class interval by the total number of readings in the data series).

The worked example given is for a data set on the river Wye in mid-Wales (see pp. 195–196). On looking at the histogram of the Wye data set (Figure 9.14) the first obvious point to note is that it is not normally distributed (i.e. it is not a classic

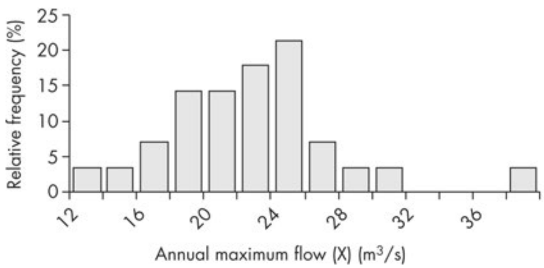


Figure 9.14 Frequency distribution of the Wye annual maximum series.

bell-shaped curve). It is important to grasp the significance of the non-normal distribution for two reasons:

1 Common statistical techniques that require normally distributed data (e.g. *t*-tests etc.) cannot be applied in flood frequency analysis.
2 It shows what you might expect: small events are more common than large floods, but that very large flood events do occur; i.e. a high magnitude, low frequency relationship.

If you were to assume that the data series is infinitely large in number and the class intervals were made extremely small, then a smooth curve can be drawn through the histogram. This is the *probability density function* which represents the smoothed version of your frequency histogram.

In flood frequency analysis there are three interrelated terms of interest. These terms are interrelated mathematically, as described in Equation 9.2 in the text below.

1 The probability of excedence: $P(X)$. This is the annual probability that a flow (Q) is greater than, or equal to, a value X. The probability is normally expressed as a unitary percentage (i.e. on a scale between 0 and 1).
2 The relative frequency: $F(X)$. This is the probability of the flow (Q) being less than a value X. This is also expressed as a unitary percentage.
3 The average recurrence interval: $T(X)$. This is expressed in years and is sometimes referred to as the return period, although this is misleading. $T(X)$ is a statistical term meaning the chance of excedence once every T years over a long record. This should not be interpreted as meaning that is exactly how many years are likely between certain size floods.
4 The annual probability of flooding: APF. This is $P(X)$ expressed as a percentage, and also the reciprocal of the average recurrence interval $T(X)$ – so the 100-year event has an APR of 1 per cent (1/100 years = 0.01 = 1 per cent). Similarly the 1 in 30 year event has an APF of 3.3 per cent.

$$P(X) = 1 - F(X)$$
$$T(X) = \frac{1}{P(X)} = \frac{1}{1 - F(X)} \qquad (9.2)$$
$$APF(\%) = \frac{1}{T(X)} \cdot 100$$

It is possible to read the values of $F(X)$ from a cumulative probability curve; this provides the simplest method of carrying out flood frequency analyses. One difficulty with using this method is that you must choose the class intervals for the histogram carefully so that the probability density function is an accurate representation of the data. Too large an interval and the distribution may be shaped incorrectly, too small and holes in the distribution will appear.

One way of avoiding the difficulties of choosing the best class interval is to use a rank order distribution. This is often referred to as a plotting position formula.

The Weibull formula

The first step in the method is to rank your annual maximum series data from low to high. In doing this you are assuming that each data point (i.e. the maximum flood event for a particular year) is independent of any others. This means that the year that the flood occurred in becomes irrelevant.

Taking the rank value, the next step is to calculate the $F(X)$ term using Equation 9.3. In this case, r refers to the rank of an individual flood event (X) within the data series and N is the total number of data points (i.e. the number of years of record):

$$F(X) = \frac{r}{N+1} \tag{9.3}$$

In applying this formula there are two important points to note:

1 The value of $F(X)$ can never reach 1 (i.e. it is asymptotic towards the value 1).
2 If you rank your data from high to low (i.e. the other way around) then you will be calculating the $P(X)$ value rather than $F(X)$. This is easily rectified by using the formula linking the two.

The worked example on pp. 188–189 gives the $F(X)$, $P(X)$ and $T(X)$ for a small catchment in mid-Wales (Table 9.3).

The Weibull formula is simple to use and effective but is not always the best description of an annual maximum series data. Some users suggest that a better fit to the data is provided by the Gringorten formula (Equation 9.4), but there are also many others:

$$F(X) = \frac{r - 0.44}{N + 0.12} \tag{9.4}$$

As illustrated in the worked example, the difference between these two formulae is not great and often the use of either one is down to personal preference.

Extrapolating beyond your data set

The probabilities derived from the Weibull and Gringorten formulae give a good description of the flood frequency within the measured stream record but do not provide enough data when you need to extrapolate beyond a known time series. This is a common hydrological problem: we need to make an estimate on the size of a flood within an average recurrence interval of 50 years but only have 25 years of streamflow record. In order to do this you need to fit a distribution to your data. There are several different ways of doing this, the method described here uses the method of moments based on the Gumbel distribution. Other distributions that are used by hydrologists include the Log-Pearson Type III and log normal. The choice of distribution is often based on personal preference and regional bias (i.e. the distribution that seems to fit flow regimes for a particular region).

Method of moments

If you assume that the data fits a Gumbel distribution then you can use the method of moments to calculate $F(X)$ values. Moments are statistical descriptors of a data set. The first moment of a data set is the mean; the second moment the standard deviation; the third moment skewness; the fourth kurtosis. To use the formulae below (Equations 9.5–9.7) you must first find the mean ($\bar{Q}$) *and standard deviation* (σ_Q) of your annual maximum data series. The symbol e in Equations 9.5–9.7 is the base number for natural logarithms or the exponential number (≈ 2.7183).

$$F(X) = e^{-e^{-b(X-a)}} \tag{9.5}$$

$$a = \bar{Q} - \frac{0.5772}{b} \tag{9.6}$$

$$b = \frac{\pi}{\sigma_Q \sqrt{6}} \tag{9.7}$$

With knowledge of $F(X)$ you can find $P(X)$ and the average recurrence interval ($T(X)$) for a certain size of flow: X. The formulae above can be rearranged to give you the size of flow that might be expected for a given average recurrence interval (Equation 9.8):

$$X = a - \frac{1}{b} \ln \ln \left(\frac{T(X)}{T(X)-1} \right) \tag{9.8}$$

In the formula above, ln represents the natural logarithm. To find the flow for a 50-year average recurrence interval you must find the natural logarithm of (50/49) and then the natural logarithm of this result.

Using this method it is possible to find the resultant flow for a given average recurrence interval that is beyond the length of your time series. The further away from the length of your time series you move the more error is likely to be involved in the estimate. As a general rule of thumb it is considered reasonable to extrapolate up to twice the length of your streamflow record, but you should not go beyond this. Thompson (1999) gives coefficients for calculating the upper and lower 90 per cent confidence limits using the lognormal or Gumbel I distribution (Table 9.3). You need to multiply your calculated flood estimates by these factors and then add or subtract these values from the calculated estimates to derive the upper and lower confidence boundaries. You will see from Table 9.3 that these are a function of the length of record you have and the recurrence interval that you are trying to estimate. So if you had 50 years of data and were trying to estimate the 100-year event, then you would be adding just over half of your original estimate onto the estimate as the upper boundary and subtracting just under 40 per cent for the lower boundary. We would be 90 per cent sure that the true flood estimate was somewhere between this broad range (see the worked example for what this looks like for the small catchment in the upper Wye).

Table 9.3 Coefficients for calculating the 90% confidence limits on annual peak discharge values estimated by the Gumbel Type I or lognormal distributions

		Recurrence interval (years)					
Years of record		*1000*	*100*	*10*	*2*	*1.1*	*1.01*
Upper	5	4.44	3.41	2.12	0.95	0.76	1.00
	10	2.11	1.65	1.07	0.58	0.57	0.76
	15	1.52	1.19	0.79	0.46	0.48	0.65
	20	1.23	0.97	0.64	0.39	0.42	0.58
	30	0.93	0.74	0.50	0.31	0.35	0.49
	40	0.77	0.61	0.42	0.27	0.31	0.43
	50	0.67	0.54	0.36	0.24	0.28	0.39
	70	0.55	0.44	0.30	0.20	0.24	0.34
	100	0.45	0.36	0.25	0.17	0.21	0.29
Lower	5	–1.22	–1.00	–0.76	–0.95	–2.12	–3.41
	10	–0.94	–0.76	–0.57	–0.58	–1.07	–1.65
	15	–0.80	–0.65	–0.48	–0.46	–0.79	–1.19
	20	–0.71	–0.58	–0.42	–0.39	–0.64	–0.97
	30	–0.60	–0.49	–0.35	–0.31	–0.50	–0.74
	40	–0.53	–0.43	–0.31	–0.27	–0.42	–0.61
	50	–0.49	–0.39	–0.28	–0.24	–0.36	–0.54
	70	–0.42	–0.34	–0.24	–0.20	–0.30	–0.44
	100	–0.37	–0.29	–0.21	–0.17	–0.25	–0.36

Source: Thompson (1999).

Worked example of flood frequency analysis

The data used to illustrate the flood frequency analysis are from the same place as the flow duration curve (the upper Wye catchment in Wales, UK). In this case it is an annual maximum series for the period 1970 until 1997 (inclusive).

In order to establish the best time of year to set a cut-off for the hydrological year, all daily mean flows above a threshold value (4.5 m^3/s) were plotted against their day number (Figure 9.15). It is clear from Figure 9.15 that high flows can occur at almost any time of the year although at the start and end of the summer (150 = 30 May; 250 = 9 September) there are slight gaps. The hydrological year from June to June is sensible to choose for this example.

The Weibull and Gringorten position plotting formulae are both applied to the data (see Table 9.4) and the $F(X)$, $P(X)$ and $T(X)$ (average recurrence interval) values calculated. The data look different from those in Figure 9.15 and from the flow duration curve because they are the peak flow values recorded in each year. This is the peak value of each storm hydrograph, which is not the same as the peak mean daily flow values.

When the Weibull and Gringorten values are plotted together (Figure 9.16) it can be seen that there is very little difference between them.

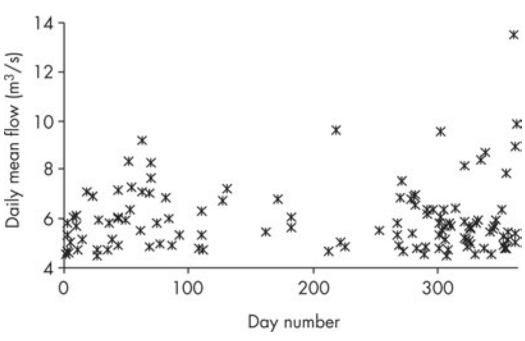


Figure 9.15 Daily mean flows above a threshold value plotted against day number (1–365) for the Wye catchment.

Table 9.4 Annual maximum series for the Wye (1970–1997) sorted using the Weibull and Gringorten position plotting formulae

Rank	*X*	*F(X) Weibull*	*F(X) Gringorten*	*P(X) Weibull*	*T(X) Weibull*
1	11.17	0.03	0.02	0.97	1.04
2	13.45	0.07	0.06	0.93	1.07
3	14.53	0.10	0.09	0.90	1.12
4	14.72	0.14	0.13	0.86	1.16
5	16.19	0.17	0.16	0.83	1.21
6	16.19	0.21	0.20	0.79	1.26
7	16.58	0.24	0.23	0.76	1.32
8	17.57	0.28	0.27	0.72	1.38
9	18.09	0.31	0.30	0.69	1.45
10	18.25	0.34	0.34	0.66	1.53
11	18.75	0.38	0.38	0.62	1.61
12	18.79	0.41	0.41	0.59	1.71
13	20.01	0.45	0.45	0.55	1.81
14	20.22	0.48	0.48	0.52	1.93
15	21.10	0.52	0.52	0.48	2.07
16	21.75	0.55	0.55	0.45	2.23
17	21.84	0.59	0.59	0.41	2.42
18	22.64	0.62	0.62	0.38	2.64
19	23.28	0.66	0.66	0.34	2.90
20	23.36	0.69	0.70	0.31	3.22
21	23.37	0.72	0.73	0.28	3.63
22	23.46	0.76	0.77	0.24	4.14
23	23.60	0.79	0.80	0.21	4.83
24	24.23	0.83	0.84	0.17	5.80
25	25.19	0.86	0.87	0.14	7.25
26	27.68	0.90	0.91	0.10	9.67
27	29.15	0.93	0.94	0.07	14.50
28	48.87	0.97	0.98	0.03	29.00

Table 9.5 Values required for the Gumbel formula, derived from the Wye data set in Table 9.4

Mean $(\bar{Q})$	*Standard deviation* (σQ)	*a value*	*b value*
21.21	6.91	18.11	0.19

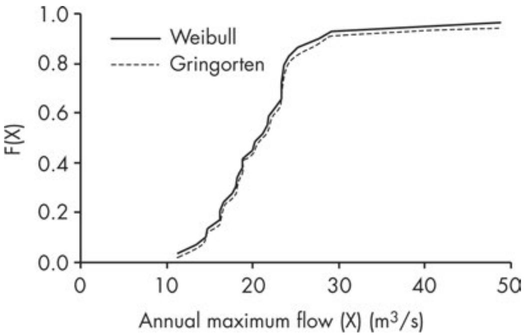


Figure 9.16 Frequency of flows less than X plotted against the X values. The $F(X)$ values are calculated using both the Weibull and Gringorten formulae.

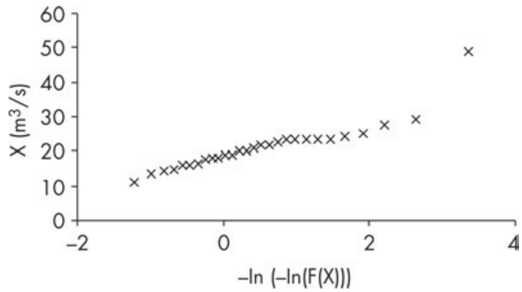


Figure 9.17 Frequency of flows less than a value X. NB The $F(X)$ values on the x-axis have undergone a transformation to fit the Gumbel distribution (see text for explanation).

When the data are plotted with a transformation to fit the Gumbel distribution they almost fit a straight line (Figure 9.17), suggesting that they do fit a distribution for extreme values such as the Gumbel but that a larger data set would be required to make an absolute straight line. A longer period of records is likely to make the extreme outlier lie further along the x-axis. The plot presented here has transformed the data to fit the Gumbel distribution. Another method of presenting this data is to plot them on Gumbel distribution paper. This provides a non-linear scale for the x-axis based on the Gumbel distribution.

Applying the method of moments and Gumbel formula to the data give some interesting results. The values used in the formula are shown below and can be easily computed. When the formula is applied to find the flow values for an average recurrence interval of 50 years it is calculated as 39.1 m^3/s. This is less than the largest flow during the record which under the Weibull formula has an average recurrence interval of 27 years. This discrepancy is due to the method of moments formula treating the highest flow as an extreme outlier. If we invert the formula we can calculate that a flood with a flow of 48.87 m^3/s (the largest on record) has an average recurrence interval of around 300 years.

Figure 9.18 shows the flood magnitude estimates for the site with the associated 90 per cent confidence limits. The estimate for the flood with a 100-year recurrence interval is therefore expressed more realistically as between 22.4 and 64.2 m^3/s at 90 per cent confidence.

Low flow frequency analysis

Where frequency analysis is concerned with low flows rather than floods, the data required are an annual minimum series. The same problem is found as for annual maximum series: which annual year to use when you have to assume that the annual minimum flows are independent of each other. At mid-latitudes in the northern hemisphere the calendar year is the most sensible, as you would expect the lowest flows to be in the summer months (i.e. the middle of the year of record). Elsewhere an analysis of when low flows occur needs to be carried out so that the hydrological year avoids splitting in the middle of a low flow period. In this case, $P(X)$ refers to the probability of an annual minimum greater than or equal to the value X. The formulae used are the same as for flood frequency analysis (Weibull etc.).

There is one major difference between flood frequency and low flow frequency analysis which has

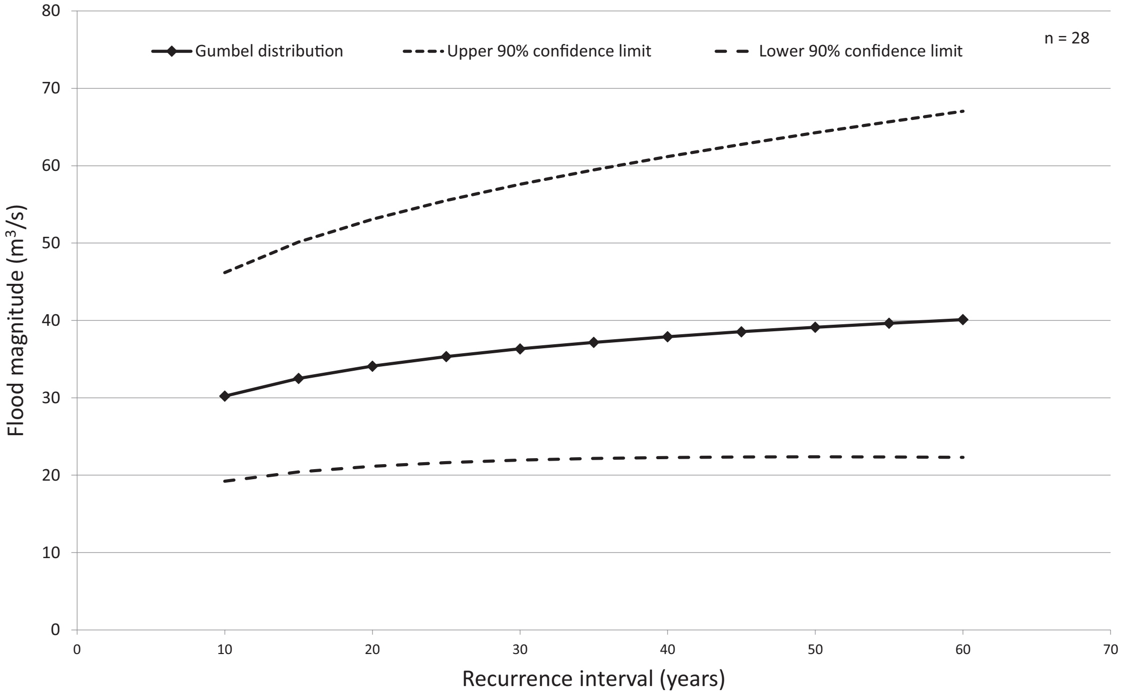


Figure 9.18 Flood magnitude estimates with the 90% confidence limits.

huge implications for the statistical methods used: there is a finite limit on how low a flow can be. In theory a flood can be of infinite size, whereas it is not possible for a low flow to be less than zero (negative flows should not exist in fresh water hydrology). This places a limit on the shape of a probability distribution, effectively truncating it on the left-hand side (see Figure 9.19).

The statistical techniques described on pp. 192–193 (for flood frequency analysis) assume a full log-normal distribution and cannot be easily applied for low flows. Another way of looking at this problem is shown in Figure 9.20 where the probabilities calculated from the Weibull formula are plotted against the annual minimum flow values. The data fit a straight line, but if we extrapolate the line further it would intersect the x-axis at a value of approximately 0.95. The implication from this is that there is a 5 per cent chance of having a flow less than zero (i.e. a negative flow). The way around this is to fit an exponential rather than a straight line to the data. This is easy to do by eye but complicated mathematically. It is beyond the level of this text

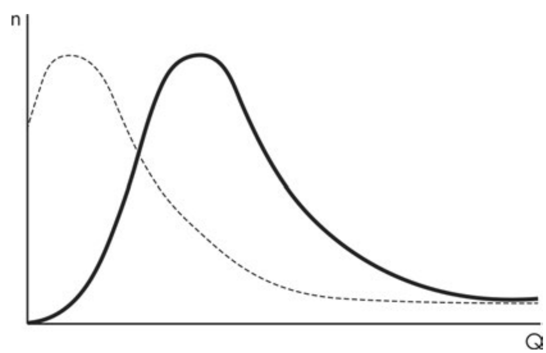


Figure 9.19 Two probability density functions. The usual log-normal distribution (solid line) is contrasted with the truncated log-normal distribution (broken line) that is possible with low flows (where the minimum flow can equal zero).

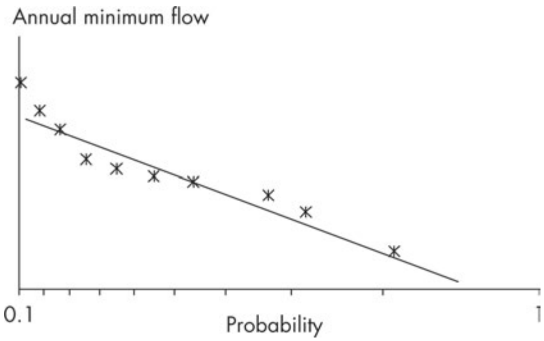


Figure 9.20 Probability values (calculated from the Weibull sorting formula) plotted on a log scale against values of annual minimum flow (hypothetical values).

to describe the technique here (see Shaw 1994, or Wang and Singh 1995 for more detail). Gordon et al. (1992) provide a simple method of overcoming this problem, without using complicated line-fitting procedures.

Limitations of frequency analysis

As with any estimation technique there are several limitations in the application of frequency analysis; three of these are major:

1 The estimation technique is only as good as the streamflow records that it is derived from. Where the records are short or of dubious quality very little of worth can be achieved through frequency analysis. As a general rule of thumb you should not extrapolate average recurrence intervals beyond twice the length of your data set, and where possible always provide confidence estimates. There is a particular problem with flood frequency analysis in that the very large floods can create problems for flow gauges and therefore this extreme data may be of dubious quality (see pp. 164–166).
2 The assumption is made that each storm or low flow event is independent of another used in the data set. This is relatively easy to guard against in annual maximum (or minimum) series, but more difficult for a peak threshold series.
3 There is an inherent assumption made that the hydrological regime has remained static during the complete period of record. This may not be true where land use, or climate change, has occurred in the catchment (see Chapter 11).

COMPUTER MODELLING IN HYDROLOGY

The easiest way of thinking about a hydrological model is to envisage streamflow as a series of numbers. Each number represents the volume of water that has flowed down the stream during a certain time period. A numerical model attempts to produce its own set of numbers, 'simulating' the flow of water down the river. There are many different ways of achieving this simulation, as will be discussed in the following section.

A model (whether mathematical, numerical or scale) is a simplification of reality. We simplify reality because the complexity of the natural world makes it difficult to understand all the processes and interactions occurring. A laboratory experiment is a similar simplification of reality; normally we are controlling all the inputs for an experiment and allowing some controlled change in a variable in order to observe the result. In constructing a computer model we are normally trying to build as good a representation of hydrological reality as we can, given our understanding of the key hydrological processes and our ability to represent these as a series of equations.

Computer modelling strategies

Black box models

The simplest forms of numerical models simulate streamflow as a direct relationship between it and another measured variable. As an example, a relationship can be derived between annual rainfall and annual runoff for a catchment (see Figure 9.21). The regression line drawn to correlate rainfall and runoff is a simulation model. If you know the annual rainfall for the catchment then you can simulate the annual runoff, using the regression relationship.

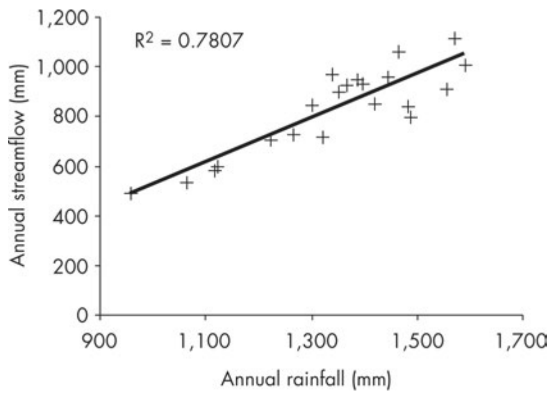


Figure 9.21 Annual rainfall vs. runoff data (1980–2000) for the Glendhu tussock catchment in the South Island of New Zealand.

Source: Data courtesy of Barry Fahey

This type of model is referred to as a black box model as it puts all the different hydrological processes that we know influence the way that water moves from rainfall to runoff into a single regression relationship. Because these types of models are based on observed data they are also called *empirical* or *metric* models. The simplicity of this type of model makes it widely applicable but its usefulness is restricted by the end-product from the model. In the example given, the regression model may be useful to estimate annual runoff in areas with the same geology and land use but it will not tell you anything about runoff at timescales less than 1 year or under different climatic and geomorphologic conditions. Another, frequently used example of a black box model is the unit hydrograph (described earlier in this chapter).

Case study

SOIL CURVE NUMBERS FOR RAINFALL–RUNOFF RELATIONSHIP

An empirical, black-box approach to predicting runoff from rainfall is the Curve Number (CN) approach developed by the United States Department of Agriculture, Soil Conservation Service (SCS 1972). The CN methodology has been used extensively in the USA for modelling rainfall–runoff relationships. The fundamental equation at the heart of the CN method is described in Equation 9.9.

$$Q=\frac{\left(P-I_a\right)^2}{P-I_a+S} \qquad (9.9)$$

Where Q is the surface runoff (mm); P is the storm precipitation total (mm); I_a is the initial abstractions (all losses before runoff begins, e.g. surface storage, rainfall interception) (mm); and S is the so-called retention parameter (mm) defined in Equation 9.10:

$$S=25.4\frac{1000}{CN}-10 \qquad (9.10)$$

Where CN refers to the curve number, which is derived using lookup tables (see SCS 1986). The CN values vary according to soil type, land use, slope and changes in antecedent soil water content. The actual number of the curve is representative of the percentage of storm rainfall that runs off as stormflow, i.e. CN of 100 corresponds to all rainfall occurring as stormflow, such as for an impervious pavement (Figure 9.22).

Empirical studies on small agricultural watersheds in the USA suggest that the initial abstraction term (I_a) can be approximated using Equation 9.11:

$$I_a = 0.2S \qquad (9.11)$$

This reduces Equation 6.9 to the form shown in Equation 9.12:

$$Q=\frac{\left(P-0.2S\right)^2}{P+0.8S} \qquad (9.12)$$

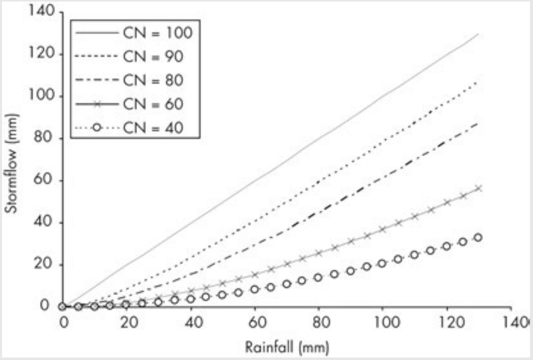


Figure 9.22 Runoff curves for a range of rainfalls.

In the CN methodology antecedent, soil moisture condition is accounted for by having three different CNs, for dry, average and wet conditions.

The CN method provides a simple solution to the problem of how to model the rainfall–runoff relationship. There are other methods to model the runoff from rainfall, e.g. the modified Green-Ampt infiltration method is frequently used in physically-based hydrological models to provide infiltration and surface runoff estimates. The simplicity provided by the CN method has many attractions but it does suffer from consequent drawbacks. The most notable of these for any analysis of land use change is that CN varies according to soil characteristics and land cover. A land use change from pasture (or suburban garden) to forest will lead to an alteration in three factors: the soil infiltration characteristics; the rainfall interception; and the antecedent soil moisture conditions. Therefore more than one factor is likely to be altered in Equations 9.9–9.12 and a simple alteration of CN may not be enough to properly account for the land use change. To fully account for this type of land use change it is necessary to use a hydrological model that explicitly accounts for antecedent soil moisture, soil infiltration characteristics and rainfall interception as distinct hydrological processes.

The CN approach has been used extensively to make runoff predictions based on a time series of rainfall (Ponce and Hawkins 1996). It has also been incorporated into more sophisticated models such as the Soil and Water Assessment Tool (SWAT; Cao et al. 2006). In this case the rainfall–runoff relationship is derived from the CN approach and combined with other hydrological processes such as evaporation estimation and river flow rates.

Lumped conceptual models

Lumped conceptual models were the first attempt to reproduce the different hydrological processes within a catchment in a numerical form. Rainfall is added to the catchment and a water budget approach used to track the losses (e.g. evaporation) and movements of water (e.g. to and from soil water storage) within the catchment area. There are many examples in the literature of lumped conceptual models used to predict streamflows (e.g. Brandt et al. 1988).

The term ‘lumped’ is used because all of the processes operate at one spatial scale – that is, they are lumped together and there is no spatial discretisation. The scale chosen is often a catchment or sometimes sub-catchments.

The term ‘conceptual’ is used because the equations governing flow rates are often deemed to be conceptually similar to the physical processes operating. So, for instance, the storage of water in a canopy or the soil may be thought of as similar to storage within a bucket. As water enters the bucket it fills up until it overflows water at a rate equal to the entry rate. At the same time it is possible to have a ‘hole’ in the bucket that allows flow out at a rate dependent on the level of water within the bucket (faster with more water). This is analogous to soil water or canopy flow but is not a detailed description such as the Darcy–Richards approximation or the Rutter model. The rate of flow through the catchment, and hence the estimated streamflow, is controlled by a series of parameters that need to be calibrated for a given catchment. Calibration is normally carried out by comparing predicted flows to measured values and adjusting (or 'optimising') the parameters until the best fit is obtained. There is considerable debate on this technique as it may sometimes be possible to obtain a similar predicted hydrograph using a completely different set of optimised parameters. It is certainly true that the optimised parameters cannot be treated as having any physical meaning and should not be transferred to catchments other than those used for calibration.

Lumped conceptual models offer a method of formulating the hydrological cycle into a water budget model that allows simulation of streamflow while also being able to 'see' the individual processes operating. This is an advance beyond black box modelling, but because the processes are represented conceptually they are sometimes referred to as grey box models (i.e. you can see partially into them). They are also referred to as parametric models.

Physically based distributed models

The rapid advances in computing power that have occurred since the 1970s mean that numerical modelling has become much easier. Freeze and Harlan (1969) were the first to formulate the idea of a numerical model that operates as a series of differential equations in a spatially distributed sense, an idea that prior to computers was unworkable. Their ideas (with some modifications from more recent research) were put into practice by several different organisations to make a physically based distributed hydrological model. Perhaps the best known of these is the Système Hydrologique Européen (SHE) model, built by a consortium of French, Danish and British organisations during the 1970s and early 1980s (Abbott et al. 1986). A model such as the SHE uses many of the process estimation techniques described in earlier chapters (e.g. Darcy's law for subsurface flow, Rutter's model for canopy interception, snow melt routines, etc.) in a water budgeting framework. Each of the equations or models used are solved for individual points within a catchment, using a grid pattern.

The principle behind this type of model is that it is totally transparent; all processes operating within a catchment are simulated as a series of physically-based equations at points distributed throughout the catchment. By 'physically-based' we mean that as far as possible the physics of the processes is being represented (rather than by empirical equations). In theory this should mean that no calibration of the model is required and spatially distributed model output for any parameter can be obtained. In reality this is far from the case. There are numerous problems associated with using a physically based, distributed model, as outlined by Beven (1989), Grayson et al. (1992) and others. The principal problem is that the amount of data required to set the initial conditions and parameterise the model is vast. The idea of obtaining saturated hydraulic conductivity measurements for every grid point in a catchment is impossible, let alone all the other parameters required. The lack of data to run the model leads to spatial averaging of parameters. There are also concerns with the size of grid used in applications (sometimes up to 1 km^2) and whether it is feasible to use the governing equations at this scale. These types of problems led Beven (1989) to query whether there really is such a thing as physically based distributed hydrological models or whether they are really just lumped conceptual models with fancier equations.

The concept of physically based distributed hydrological modelling is noble, but in reality the models have not produced the results that might have been expected. They are certainly unwieldy to use and have many simplifications that make the terminology doubtful. However they have been useful for gaining a greater understanding of our knowledge base in hydrological processes. The approach taken, with its lack of reliance on calibration, still offers the only way of investigating issues of land use change and predicting flows in ungauged catchments.

Hydrological modelling for specific needs

In many cases where streamflow needs to be estimated, the use of a physically based model is akin to using the proverbial 'sledgehammer to crack a walnut'. With the continuing increase in computing power there are numerous tools available to the hydrologist to build their own computer model to simulate a particular situation of interest. These tools range from Geographic Information Systems (GIS) with attached dynamic modelling languages to object-oriented languages that can use icon-linked modelling approaches (e.g. McKim et al. 1993). This perhaps offers a future role for hydrological modelling away from the large modelling packages such as SHE. In essence it allows the hydrologist to simulate streamflow based on a detailed knowledge of catchment processes of importance for the particular region of interest.

FLOW ASSESSMENT FOR STREAM ECOLOGY

Managers of river systems frequently need information on the amount of flow required to sustain the current river ecology. This is in order to ascertain how much water is available for out of stream usage (e.g. irrigation) without placing detrimental stress on the current river ecology. Consequently a branch of science has been developed that combines knowledge of river hydraulics with aquatic ecology to provide this information. This is described in the following section. The discussion is continued further in Chapter 11 where there is a description of water allocation methodology.

Jowett (1997) divides the methodology used for instream flow assessment into three types with an increasing complexity: *historic flow*, *hydraulic* and *habitat methods*. The historic approach sets water abstraction limits based on a historical flow range; the hydraulic method uses the relationship between hydraulic parameters (e.g. wetted perimeter, stream depth, etc.) and stream health; and the habitat method uses actual measurements of stream health with changes in flow regime to predict the impacts of flow changes. The way these methods treat the relationship between increased streamflow and the biological response is shown in Figure 9.23. The historic method assumes that there is a linear relationship so that more flow results in a greater biological response. The hydraulic method recognises that stream beds are non-linear in form and therefore a small change in flow may result in large increases in biological productivity but that this decreases as the flow increases. The habitat method recognises that there is a maxima in the biological productivity and high flows may lead to decreasing biological response.

Case study

WATYIELD – MODELLING CHANGES IN WATER YIELD FROM ALTERING LAND COVER CHANGE

The degree of vegetation cover in a catchment will affect the amount of water flowing down a stream. The physical processes that cause this effect have been described in Chapter 3 and the impacts of this change are discussed in Chapter 11. A water balance model has been developed to quantify the impact of land use changes on the stream discharge. The model is simple to use and can be downloaded for free from the World Wide Web (look for WATYIELD at http://icm.landcare research.co.nz). Also available at this site are a series of reports that help parameterise the model and a full user's guide.

The WATYIELD model was developed for New Zealand conditions and works best in a humid temperate environment. It has been designed for catchments up to around 50 km^2 in size. In the modelling terminology outlined earlier in this chapter, WATYIELD could be described as a lumped, conceptual model. However, there is detailed process representation of rainfall interception and soil moisture storage within the model so it moves slightly towards being physically based. The spatial representation is at the catchment scale; although it is possible to split a catchment into sub-sections with different vegetation covers and rainfall distributions. However, these sub-sections have no spatial differentiation within the catchment, i.e. the model doesn't know where they are within a catchment, just that there are subsections. For catchments larger than 50 km^2 the underlying assumptions of spatial uniformity start to break down and it is necessary to start introducing elements such as flow routing down a stream (presently ignored at the daily time step of WATYIELD).

WATYIELD works by adding daily rainfall to two storage terms which release water to a river based on hydrograph recession coefficients (a full description of the model can be found in Fahey et al. 2004). The storage terms represent soil moisture and a deeper groundwater store. Daily rainfall is processed by the model so that any interception loss from a vegetation canopy is removed and all the resultant rainfall infiltrates into the soil moisture store. In order to operate the model a daily rainfall record, potential evapotranspiration, soil parameters and knowledge about flow characteristics from a nearby stream are required. Much of this type of data is readily available from the scientific literature and resource management databases.

WATYIELD has been applied to a 23 km^2 catchment (Rocky Gully) in the South Island of New Zealand to investigate two possible land use change scenarios. The current vegetation cover for the catchment is a mixture of tall tussock grassland, pasture grasses and a small amount of scrubland forest. The catchment has an altitude range from 580 m to 1,350 m with an increasing rainfall with altitude.

In testing the model against daily streamflow from 1989–2001, WATYIELD was able to predict mean annual flow within 2 per cent accuracy and mean annual 7-day low flow within 3 per cent. The two scenarios simulated were:

- 40 per cent of the catchment was converted to plantation forestry (*Pinus radiata*). All the planting occurred in the lower half of the catchment;
- 50 per cent of the catchment was converted from tussock to pasture grassland. All the tussock grassland in the upper half of the catchment was replaced with pasture species.

Each of these is a realistic land use change scenario for the region; conversion of pasture land to forestry is common practice, as is 'improving' grassland by over-sowing with rye grass species. The land use change scenarios were simulated in the model using the 1989–2001 rainfall data (i.e. repeating the earlier simulation but with a different land cover).

The results from the modelling are shown in Table 9.6. An initial look at the results suggests a surprising result: the amount of interception loss from a 40 per cent increase in forestry does not transfer through into much of a change in mean annual streamflow or low flows. There is a larger change in flow regime from the replacement of tussock grassland in the upper catchment; despite this land use change resulting in a lowering of interception loss (tussock grassland has higher interception losses than pasture grass). The reason for these results is that it is the upper part of the catchment, with a higher rainfall, that produces most of the streamflow, particularly the low flows. Hence a change in land use in the lower section makes a relatively small change in the flow regime. However, a change in land use in the upper region of the catchment has a larger effect

Table 9.6 Results from WATYIELD modelling of land use change

Flow measure	*Scenario 1 (forestry in lower half replaces pasture)*	*Scenario 2 (replacement of tussock grassland with pasture)*
Mean annual flow	Reduced by 6%	Reduced by 7%
Mean annual 7-day low flow	Reduced by 3%	Reduced by 7%

because this is where the effective rainfall is occurring. A change from tussock to pasture grassland increases the transpiration loss which more than offsets the decrease in canopy interception.

In this case WATYIELD was able to tease out the difference between canopy interception and canopy transpiration. The difference between the two is what made the most difference in the simulations. The final scenario modelled was to place the forestry in the upper reaches of the catchment; this reduced flows by around 25 per cent. However this is a highly unlikely land use change scenario as commercial forestry at this latitude does not normally extend beyond 850 m above sea level.

The most well-known of the *historic flow methods* is 'Montana method' proposed by Tennant (1976), also called the Tennant method. Tennant (1976) used hydraulic data from 11 streams in the USA and knowledge about depths and velocities required to sustain aquatic life to suggest that 10 per cent of average flow is the lower limit for aquatic life. Tennant (1976) also recommended that 30 per cent of average flow provides a satisfactory stream environment. With relatively easily derived flow information (i.e. average flow) a new flow regime can be set for a river that takes into account the instream values. However this approach precludes the possibility that a stream could be enhanced by a non-natural flow regime. This is especially true where there is an upstream reservoir, in which case flows can be manipulated to improve the aquatic environment, not just maintain what is presently there.

The *hydraulic method* requires measurements of hydraulic data such as wetted perimeter, width, velocity and depth at a series of cross sections. Then, using either rating curves (i.e. the stage–discharge relationship described in Chapter 8) or an equation such as Manning's (see Chapter 8), the variations in a hydraulic parameter with flow can be derived. The most commonly used hydraulic parameter is the wetted perimeter because it takes into account the area of streambed where periphyton and invertebrates live. A healthy periphyton and invertebrate community generally leads to a healthy river ecosystem. The variation in wetted perimeter with flow is drawn in the same way as represented by the broken line in Figure 9.23. The minimum flow for river is normally defined by where the hydraulic parameter (e.g. wetted perimeter) starts to decline sharply to zero. The hydraulic method has the advantage over the historic method that it

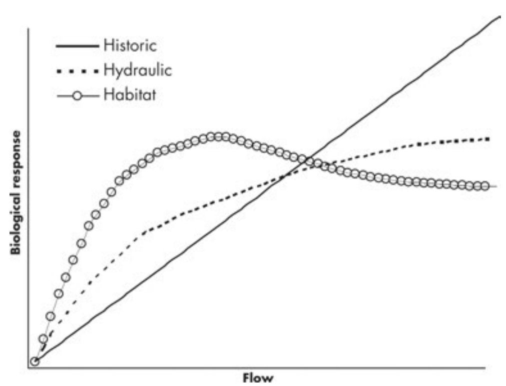


Figure 9.23 Hypothetical relationships showing biological response to increasing streamflow as modelled by historic, hydraulic and habitat methods.

Source: Adapted from Jowett (1997)

takes into account the actual streambed morphology which may differ markedly between rivers.

The *habitat method* extends the hydraulic method by taking the hydraulic information and combining it with knowledge of how different aquatic species survive in those flow regimes. In this way the appropriate flow regime can be designed with particular aquatic species in mind. In the case of fish, some prefer shallow turbulent streams compared to deep, slow moving rivers. The habitat method allows differentiation between these so that a flow regime can be set with protection, or enhancement, of a particular species in mind.

The most common use of the habitat method is the Instream Flow Incremental Methodology (IFIM; Irvine et al. 1987; Navarro et al. 1994) which has been developed into computer models such as PHABSIM (Physical HABitat SIMulation; Milhous et al. 1989; Gallagher and Gard 1999) and RHYHABSIM (River HYdraulic HABitat SIMulation; Jowett 1997).

The habitat method focuses on a particular species and life stage at a time, and investigates its response at a particular flow. For each cell in a two-dimensional grid, velocity, depth, substrate and possibly other parameters (e.g. cover) at the given flow are converted into suitability values, one for each parameter. These suitability values are combined (usually multiplied) and multiplied by the cell area to give an area of usable habitat (also called weighted usable area, WUA). Finally, all the usable habitat cell areas are summed to give a total habitat area (total WUA) for the reach at the given flow. The whole procedure is repeated for other flows until a graph of usable habitat area versus flow for the given species has been produced. This graph has a typical shape, as shown in Figure 9.22, with a rising part, a maximum and a decline. The decline occurs when the velocity and/or depth exceed those preferred by the given species and life stage. In large rivers, the curve may predict that physical habitat will be at a maximum at less than naturally occurring flows (Jowett 1997).

Models such as PHABSIM have been used successfully in many places around the world to advise water managers on the best flow regime for particular aquatic species. It should be noted that this is a physical approach to the problem, i.e. it takes account of the physical flow regime of the river with no consideration of water quality parameters. The assumption is made that water quality does not change with the proposed changing flow regime.

ESSAY QUESTIONS

1. **Find a scientific paper in the literature that uses a hydrological model and evaluate the type of model and its strengths and weaknesses for the study concerned.**
2. **Outline the limitations of the unit hydrograph when used as a predictive tool and attempt to explain its success despite these limitations.**
3. **Describe the types of information that can be derived from a flow duration curve and explain the use of that information in hydrology.**
4. **Explain why interpretation of flood (or low flow) frequency analysis may be fraught with difficulty.**
5. **Describe the data (and measurement equipment) required for using the IFIM approach to look at the habitat requirements for a particular aquatic species.**

FURTHER READING

Beven, K. (2012) *Rainfall-runoff modelling: The primer* (2nd edition). Wiley-Blackwell, Chichester.
An introduction to modelling in hydrology.

Callow, P. and Petts G.E. (eds) (1994) *The rivers handbook: Hydrological and ecological principles*. Blackwell Publishing, Hoboken, N.J.
Gives further detail on combining hydraulic, hydrological and ecological principles for river management.

Dingman, S.L. (2014) *Physical hydrology* (3rd edition). Waveland Press, Long Grove, I.L.
A high level text with good detail on analytical techniques.

Maidment, D.R. (ed.) (1993) *Handbook of hydrology*. McGraw-Hill, New York.
An older text, but with useful sections. Part 3 concentrates on hydrological analysis techniques.

Shaw, E.M., Beven, K.J., Chappell, N.A. and Lamb, R. (2011) *Hydrology in practice* (4th edition). Spon Press, London.
An engineering text with good detail on analytical techniques.

10

WATER QUALITY

LEARNING OBJECTIVES

When you have finished reading this chapter you should have:

- An understanding of water quality as an issue in hydrology and how it ties into water quantity.
- A knowledge of the main parameters used to assess water quality and what affects their levels in waterbodies.
- A knowledge of the measurement techniques and sampling methodology for assessing water quality.
- A knowledge of techniques used to control water pollution and manage water quality.

This chapter identifies the different types of pollutants that can be found in aquatic environments and describes their major sources, especially where elevated levels may be found and what impact their presence has on aquatic ecology. The chapter also outlines the methods used to measure water quality parameters. This is followed by a description of the management techniques used to control water quality in a river catchment.

Traditionally, hydrology has been interested purely in the amount of water in a particular area: water quantity. This is frequently referred to as physical hydrology. If, however, we take a wider remit for hydrology – to include the availability of water for human consumption – then issues of water quality are of equal importance to quantity. There are three strong arguments as to why hydrology should consider water quality an area worthy of study.

1 *The interlink between water quality and quantity*. Many water quality issues are directly linked to the amount of water available for dilution and dispersion of pollutants, whether they be natural or anthropogenic in source. It is virtually impossible to study one without the other. An example of this is shown in the Case Study of the River Thames through London (pp. 209–211).
2 *The interlink between hydrological processes and water quality*. The method by which pollutants transfer from the land into the aquatic environment

is intrinsically linked with the hydrological pathway (i.e. the route by which the water moves from precipitation into a stream), and hence the hydrological processes occurring. A good example of this is in Heppell et al. (1999) where the mechanisms of herbicide transport from field to stream are linked to runoff pathways in a clay catchment.

3 *Employment of hydrologists*. It is rare for someone employed in water resource management to be entirely concerned with water quantity, with no regard for quality issues. The maintenance of water quality is not just for drinking water (traditionally an engineer's role) but at a wider scale can be for maintaining the **amenity value** of rivers and streams.

It is easy to think of water quality purely in terms of pollution; i.e. waste substances entering a river system as a result of human activity. This is an important issue in water quality analysis but is by no means the only one. One of the largest water quality issues is the amount of suspended sediment in a river, which is frequently a completely natural process. Suspended sediment has severe implications for the drinking water quality of a river, but also for other hydrological concerns such as ecological impacts and implications for reservoir design. As soon as a river is dammed the water velocity will slow down. Simple knowledge of the **Hjulstrom curve** (see Figure 10.1) tells us that this will result in the deposition of suspended sediment. That deposition will eventually reduce the capacity of the reservoir held behind the dam. In high-energy river systems, for even a very large reservoir, a dramatic reduction in capacity can take place within two to four decades. It is critically important for a hydrologist involved in reservoir design to have some feeling for the quantities of suspended sediment so that the lifespan of a reservoir can be calculated. In South Korea, reservoir management includes understanding the sediment plume entering a reservoir during the rainy season and using a multiple level abstraction to release this sediment laden water during the wet season, i.e. minimising sedimentation in the dam (Kim et al. 2007).

Spatial variations in water quality may be influenced by many different environmental factors (e.g. climate, geology, weathering processes, vegetation cover and anthropogenic influences). Often it is a combination of these factors that makes a particular water quality issue salient for a particular area. An example of this is acid rain (also discussed in Chapter 2) as a particular problem for north-eastern North America and Scandinavia. The sources of the acid rain are fossil fuel burning power stations and industry. It becomes a particular problem in these areas for a number of reasons: it is close to the sources of acid rain; high rainfall contributes a lot of acid to the soil; the soils are thin after extensive glaciation and derived from very old rocks; and the soils are heavily leached (have had a lot of water passing through them over a long time period) and have a low buffering capacity (see p. 47). This combination of influences means that the water in the rivers has a low **pH**, and – of particular concern to gill-bearing aquatic fauna – has a high dissolved aluminium content because acidic water leaches aluminum out of the soil.

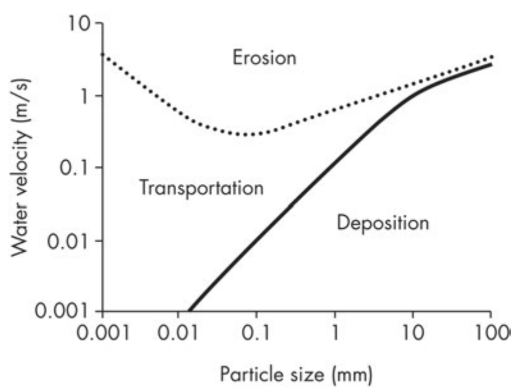


Figure 10.1 The Hjulstrom curve relating stream velocity to the erosion/deposition characteristics for different sized particles (x-axis). In general, the slower the water moves, the finer the particles that are deposited, and the faster the water moves the larger the particles being transported.

Having argued for the role of natural water-quality issues to be considered seriously, the reader will find that the majority of this chapter deals with human-induced water-quality issues. This is an inevitable response to the world we live in where we place huge pressures on the river systems as repositories of waste products. It is also important to study these issues because they are something that humans can have some control over, unlike many natural water-quality issues.

Case study

THE RIVER THAMES THROUGH LONDON: WATER QUALITY CHANGE

The River Thames as it flows through London is one of the great tourist sights of Europe. It is an integral part of London, not just for its scenic attraction but also as a transport route right into the heart of a modern thriving city. The river also has a large part to play in London's water resources, both as a source of drinking water and a disposal site for waste.

London has a long history of water-quality problems on the Thames, but it has not always been so. Prior to the nineteenth century, domestic waste from London was collected in cesspools and then used as fertiliser on agricultural land (hence the use of the term 'sewage farm' for sewage treatment stations). The Thames maintained a fish population, and salmon from the river were sold for general consumption. With the introduction of compulsory water closets (i.e. flushing toilets) in 1843 and the rise in factory waste during the Industrial Revolution, things started to change dramatically for the worse during the nineteenth century. The majority of London's waste went through poorly constructed sewers (often leaking into shallow aquifers which supplied drinking water) straight into the Thames without any form of treatment. In 1854 there was an outbreak of cholera in London that resulted in up to 10,000 deaths. In a famous epidemiological study Dr John Snow was able to link the cholera to sewage pollution in water drawn from shallow aquifers. The culmination of this was 'the year of the great stink' in 1856. The smell of untreated waste in the Thames was so bad that disinfected sheets had to be hung from windows in the Houses of Parliament to lessen discomfort for the lawmakers of the day. In the best NIMBY ('not in my back yard') tradition this spurred parliament into action and in the following decade, radical changes were made to the way that London used the River Thames. Water abstraction for drinking was only permitted upstream of tidal limits and London's sewage was piped downstream to Beckton where it was discharged (still untreated) into the Thames on an ebb tide.

The result of these reforms was a radical improvement of the river water quality through central London; but there was still a major problem downstream of Beckton. The improvements were not to last, however, as by the middle of the twentieth century the Thames was effectively a dead river (i.e. sustained no fish population and had a dissolved oxygen concentration of zero for long periods during the summer). This was the result of several factors: a rapidly increasing population, increasing industrialisation, a lack of investment in sewage treatment and bomb damage during the Second World War.

Since the 1950s the Thames has been steadily improving. Now there is a resident fish population and migratory salmon can move up the Thames. This improvement has been achieved through an upgrading of the many sewage treatment works that discharge into the Thames and its tributaries. The Environment Agency (the statutory authority responsible for rivers in England) has much to do with the management of the lower Thames and proudly proclaims that the Thames 'is one of the cleanest metropolitan rivers in the world'. How realistic is this claim?

There is no doubt that the Thames has been transformed remarkably from the 'dead' river of 60 years ago into something far cleaner, but there are two problems remaining for the management of the Thames through London, and for one of these nothing can be done.

- The Thames is a relatively small river that does not have the flushing potential of other large rivers; therefore it cannot cleanse itself very easily.
- The sewer network underneath London has not been designed for a large modern city and cannot cope with the strains put on it.

At Westminster (in front of the Houses of Parliament) the Thames is over 300 m wide; this is confined from the width of 800 m evident during Roman times. This great width belies a relatively small flow of fresh water. It appears much larger than in reality because of its use for navigation and the tidal influence. The average flow rate for the Thames is 53 cumecs, rising to 130 cumecs under high flows. In Table 10.1 this is compared with rivers that flow through other major cities. In Seoul, a similarly sized capital city, the Han River is over seven times larger than the Thames. The effect of the small flow in the Thames is that it does not have great flushing power. During the summer months it may take a body of water three months to move from west London to the open sea. On each tide it may move up to 14 km in total but this results in less than a kilometre movement downstream. If this body of water is polluted in some way then it is not receiving much dilution or dispersion during the long trip through London.

The second important factor is the poor state of London's sewers. Prior to Sir Joseph Bazalgette's sewer network of 1864 the old tributaries of the Thames acted as sewers, taking waste water directly to the Thames. Bazalgette's grand sewerage scheme intercepted these rivers and transported the sewage through a large pipe to east London. This system still exists today. The actual sewerage network is very well built and still works effectively. The problem is that it is unable to cope with the volume of waste expected to travel through it, particularly when it rains, as it is a combined stormwater–sewage network. The original tributaries of the Thames, such as the Fleet, still exist under London and any storm runoff is channelled into them. When the volume of stormwater and sewage is too great for the sewers, the rivers act as overspills and take the untreated sewage directly into the Thames. This is a particular problem during summer storms when the volume of water flowing down the Thames is low and cannot dilute the waste effectively.

Table 10.1 Comparison of rivers flowing through major cities

River	*Mean annual flow (m^3/s)*	*City on river or estuary*	*Population in metropolitan area (million)*
Thames	82	London	12.0
Seine	268	Paris	9.93
Hudson	387	New York	19.3
Han	615	Seoul	10.3
Rhine	2,219	Rotterdam	1.1
Paraná/Uruguay	22,000	Buenos Aires	11.6

Source: Flow data from Global Runoff Data Centre

To combat this problem Thames Water Utilities (part of the private company that treats London's sewage) operate two boats especially designed to inject oxygen directly into the water. These boats can float with a body of sewage-polluted water, injecting oxygen so that the dissolved oxygen level does not reach levels that would be harmful to fish and other aquatic creatures. To help in the tracking of a polluted body of water there are water-quality monitoring stations attached to bridges over the Thames. These stations measure temperature, dissolved oxygen concentration and electrical conductivity at 15-minute intervals and are monitored by the Environment Agency as they are received in real time at the London office.

In addition to the oxygen-injecting boats, there is tight water-quality management for the River Thames through London. This is operated by the Thames Estuary Partnership (http://thamesestuarypartnership.org), a group of interested bodies including the Environment Agency. Their remit includes other factors such as protecting London from flooding (using the Thames Barrier), but also setting higher effluent standards for sewage treatment works during the summer. The emphasis is on flexibility in their management of the Thames. In 2016, construction started on the Thames Tideway Tunnel – a 25 km interception, storage and transfer tunnel running below the Thames which will take waste from London to Beckton Sewage Treatment Works. This major upgrade of London's sewerage system will significantly reduce overflows of untreated sewerage into the river Thames. There is no question that the River Thames has improved from 50 years ago. In many respects it is a river transformed, but it still has major water-quality problems such as you would expect to find where a small river is the receptacle for the treated waste of over 10 million people. The water-quality management of a river like the Thames needs consideration of many facets of hydrology: understanding pollutants, knowledge of stormflow peaks from large rainfall events, and streamflow statistics.

Before looking at the water-quality issues of substances within a river system it is worth considering how they reach water bodies. In studying water pollution it is traditional to differentiate between *point source* and *diffuse* pollutants. As the terminology suggests, point sources are discrete places in space (e.g. a sewage treatment works) where pollutants originate. Diffuse sources are spread over a much greater land area and the exact locations cannot be specified. Examples of diffuse pollution are excess fertilisers and pesticides from agricultural production. The splitting of pollutants into diffuse and point sources has some merit for designing preventative strategies but like most categorisations there are considerable overlaps. Although a sewage treatment works can be thought of as a point source when it discharges effluent into a stream, it has actually gathered its sewage from a large diffuse area. If there is a particular problem with a sewage treatment works effluent, it may be a result of accumulated diffuse source pollution rather than the actual sewage treatment works itself.

A more useful categorisation of water pollutants is to look at their impacts on the river system. In this way we can differentiate between three major types of pollutants.

- *Toxic compounds*, which cause damage to biological activity in the aquatic environment.
- *Oxygen balance affecting compounds*, which either consume oxygen or inhibit the transfer of oxygen between air and water. This would also include thermal pollution as warm water does not hold as much dissolved oxygen as cold water (see pp. 215–216).
- *Suspended solids* – inert solid particles suspended in the water.

Whether we approve or not, rivers are receptacles for large amounts of waste produced by humans. Frequently this is deliberate and is due to the ability of rivers to cope with waste through degradation, dilution and dispersion. Just how quickly these three processes operate is dependent on the pollutant load already present in the river, the temperature and pH of the water, the amount of water flowing down the river and the mixing potential of the river. The last two of these are river flow characteristics that will in turn be influenced by the time of year, the nature of flow in the river (e.g. the shape of the flow duration curve), and the velocity and turbulence of flow. This demonstrates the strong interrelationship that exists between water quality and water quantity in a river system.

One remarkable feature about rivers is that given enough time and a reasonable pollution loading, rivers will recover from the input of many pollutant types. That is not to say that considerable harm cannot be done through water pollution incidents, but by and large the river system will recover so long as the pollution loading is temporary. An example of this can be seen in the **oxygen sag curve** (see Figure 10.2) that is commonly seen below point sources of organic pollution (e.g. sewage effluent). The curve shows that upon entering the river there is an instant drop in dissolved oxygen content. This is caused by bacteria and other micro-organisms in the river feeding on the organic matter in the stream and using any available dissolved oxygen. This would have a severe impact on any aquatic fauna unable to move away from this zone of low dissolved oxygen. As the pollutant load moves downstream the degradation, dilution and dispersal starts to take effect and oxygen levels start to recover in the river. The shape of the curve, especially the distance downstream until recovery, is highly dependent on the flow regime of the receiving river. A fast flowing, readily oxygenated stream will recover much faster than a slow-moving river. Large rivers will have a faster recovery time (and the depth of sag will be less) than small streams, due to the amount of dilution occurring.

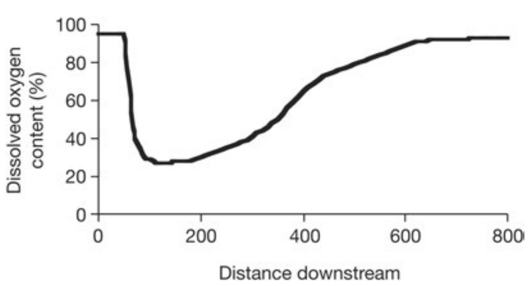


Figure 10.2 Hypothetical dissolved oxygen sag curve. The point at which the curve first sags is the point source of an organic pollutant. The distance downstream has no units attached as it will depend on the size of the river.

WATER-QUALITY PARAMETERS

To analyse the water quality within a river, consideration has to be given to what type of test may be carried out and the sampling pattern to be used. There are numerous parameters that can be measured, and each is important for the part they play in an overall water-quality story. It is not necessary to measure them all for a single water-quality analysis study; instead the relevant parameters for a particular study should be identified. This can be done using a priori knowledge of the water-quality issues being studied. To aid in this, different parameters are discussed here with respect to their source; what type of levels might be expected in natural rivers; and the impact they have on a river ecosystem. However, a critical point to remember about any kind of water quality measurement is that it is just a 'snap shot' in time; water quality parameters can vary markedly over time.

The first distinction that can be made is between physical and chemical parameters. With chemical parameters it is the concentration of a particular chemical substance that is being assessed. With physical parameters it is a physical measurement being made, normally measuring the amount of something within a water sample.

Physical parameters

Temperature

The temperature of water in a river is an important consideration for several reasons. The most important feature of temperature is the interdependence it has with dissolved oxygen content (see pp. 215–216). Warm water holds less dissolved oxygen than colder water. The dissolved oxygen content is critical in allowing aquatic fauna to respire, so temperature is also indirectly important in this manner. Water temperature is also a controlling factor in the rate of chemical reactions occurring within a river. Warm water will increase the rate of many chemical reactions occurring in a river, and it is able to dissolve more substances. This is due to a weakening of the hydrogen bonds and a greater ability of the bipolar molecules to surround anions and cations.

Warm water may enter a river as thermal pollution from power stations and other industrial processes. In many power stations (gas, coal and nuclear) water is used as a coolant in addition to the generation of steam to drive turbines. Because of this, power stations are frequently located near a river or lake to provide the water source. It is normal for the power stations to have procedures in place so that hot water is not discharged directly into a river; however, despite the cooling processes used, the water is frequently 1–2 °C warmer on discharge. The impact that this has on a river system will be dependent on the river size (i.e. degree of dilution and rate of dispersion).

Dissolved solids

In the first chapter, the remarkable ability of water to act as a solvent was described. As water passes through a soil column or over a soil surface it will dissolve many substances attached to the soil particles. Equally water will dissolve particles from the air as it passes through the atmosphere as rain. The amount of dissolved substances in a water sample is referred to as the **total dissolved solids (TDS)**. The higher the level of TDS the more contaminated a water body may be, whether that be from natural or anthropogenic sources. Meybeck (1981) estimates that the global average TDS load in rivers is around 100 mg/l, but it may rise considerably higher (e.g. the Colorado River has an average TDS of 703 mg/l).

Electrical conductivity

The ability of a water sample to transmit electrical current (its conductivity) is directly proportional to the concentration of dissolved ions. Pure, distilled water will still conduct electricity but the more dissolved ions in water the higher its electrical conductivity. This means that there is a linear relationship between TDS and conductivity which can be derived (Equation 10.1).

$$K = \frac{\text{Conductivity}}{\text{TDS}} \text{ or } \text{TDS} = \frac{\text{Conductivity}}{K} \qquad (10.1)$$

This relationship gives a very good surrogate measure for TDS. The K term is a constant (usually between 0.55 and 0.75) that can be estimated by taking several measurements of conductivity with differing TDS levels. Conductivity is a simple measurement to take as there are many robust field instruments that will give an instant reading. This can then be related to the TDS level at a later stage. Electrical conductivity is measured in Siemens per metre, although the usual expression is microsiemens per centimetre (μS/cm). Rivers normally have a conductivity between 10 and 1,000 μS/cm.

Suspended solids

The amount of suspended solids has been highlighted at the start of this chapter as a key measure of water quality. The carrying of suspended sediment in a river is part of the natural erosion and sediment transport process. The sediment will be deposited at any stage when the river velocity drops and conversely it will be picked up again with higher river velocities (see Figure 10.1). In this manner the **total suspended solids (TSS)** load will vary in space and time. The amount of TSS in a river will affect the aquatic fauna, because it is difficult for egg-laying fish and invertebrates to breed in an environment of high sediment. Many species of fish, for example, need clean, well-oxygenated gravels in which to spawn. Suspended sediment is frequently inert, as in the case of most clay and silt particles, but it can be organic in content and therefore have an oxygen demand.

TSS is expressed in mg/l for a water sample but frequently uses other units when describing sediment

Table 10.2 Sediment discharge, total river discharge (averaged over several years) and average total suspended solids (TSS) for selected large river systems

River (country)	*Sediment discharge (10^3 tonnes/yr)*	*Discharge (km^3/yr)*	*Average TSS (mg/l)*
Zaïre (Zaïre)	43,000	1,250	0.03
Amazon (Brazil)	900,000	6,300	0.14
Danube (Romania)	67,000	206	0.33
Mississippi (USA)	210,000	580	0.36
Murray (Australia)	30,000	22	1.36
Ganges-Brahmaputra (Bangladesh)	1,670,000	971	1.72
Huanghe or Yellow (China)	1,080,000	49	22.04

Source: Data from Milliman and Meade (1983)

load. Table 10.2 shows some values of sediment discharge (annual totals) and calculates an average TSS from the data. It is remarkable to see the data in this form, enabling contrast to be drawn between the different rivers. Although the Amazon delivers a huge amount of sediment to the oceans it has a relatively low average TSS, a reflection of the extremely high discharge. In contrast to this the Huanghe river (sometimes referred to as the Yellow river due to the high sediment load) is virtually a soup! It must be noted that these are average values over a year and that the TSS will vary considerably during an annual cycle (the TSS will rise considerably during a flood).

Turbidity

Turbidity is a measure of the cloudiness of water. The cloudiness is caused by suspended solids and gas bubbles within the water sample, so TSS and turbidity are directly related. Turbidity is measured as the amount of light scattered by the suspended particles in the water. A beam of light of known luminosity is shone through a sample and the amount reaching the other side is measured. This is compared to a standard solution of formazin. The units for turbidity are either FTU (formazin turbidity units) or NTU (normalised turbidity units); they are identical. Turbidity is a critical measure of water quality for the same reasons as TSS. It is a simpler measurement to make, especially in the field, and therefore it is sometimes used as a surrogate for TSS.

Chemical parameters

pH

Chemists think of water as naturally disassociating into two separate ions: the hydroxide (OH^-) and hydrogen (H^+) ions.

$$H_2O \rightleftharpoons OH^- + H^+$$

The acidity of water is given by the hydrogen ion, and hence pH (the measure of acidity) is a measure of the concentration of hydrogen ions present. In fact it is the log of the inverse concentration of hydrogen ions (Equation 10.2).

$$P^H = \log \frac{1}{\left[H^+\right]} \tag{10.2}$$

This works out on a scale between 1 and 14, with 7 being neutral. A pH value less than 7 indicates an acid solution; greater than 7 a basic solution (also called alkaline). It is important to bear in mind that because the pH scale is logarithmic (base 10) a solution with pH value 5 is ten times as acidic as one with pH value 6.

In natural waters the pH level may vary considerably. Rainwater will naturally have a pH value less than 7, due to the absorption of gases such as carbon dioxide by the rainwater. This forms a weak carbonic acid, increasing the concentration of hydrogen ions in solution. The normal pH of rainfall is somewhere between 5 and 6 but may drop as low as 4, particularly if there is industrial air pollution nearby. For example, Zhao and Sun (1986) report a pH value of 4.02 in Guiyang city, China, during 1982.

Acidic substances may also be absorbed easily as water passes through a soil column. A particular example of this is water derived from peat, which will absorb organic substances. These form organic acids, giving peat-derived water a brown tinge and a low pH value. At the other end of the spectrum rivers that drain carbonate-rich rocks (e.g. limestone and chalk), have a higher pH due to the dissolved bicarbonate ions.

The pH value of rivers is important for the aquatic fauna living within them. The acidity of a river is an important control for the amount of dissolved ions present, particularly metal species. The more acidic a river is the more metallic ions will be held in solution. For fish it is often the level of dissolved aluminium that is critical for their survival in low pH waters. The aluminium is derived from the breakdown of alumino-silicate minerals in clay, a process that is enhanced by acidic water. Water with a pH between 6 and 9 is unlikely to be harmful to fish. Once it drops below 6 it becomes harmful for breeding, and salmonid species (e.g. trout and salmon) cannot survive at a pH lower than 4. Equally a pH higher than 10 is toxic to most fish species (Alabaster and Lloyd 1980). Table 10.3 summarises the effect of decreasing pH (i.e. increasing acidity) on aquatic ecology.

Mention needs to be made of the confusing terminology regarding **alkalinity**. Alkalinity is a measure of the capacity to absorb hydrogen ions without a change in pH (Viessman and Hammer 1998). This is influenced by the concentration of hydroxide, bicarbonate or carbonate ions. In water-quality analysis the term 'alkalinity' is used almost exclusively to refer to the concentration of bicarbonate (HCO^{3-}) ions because this is the most variable of the three. The bicarbonate ions are derived from the percolation of water through calcareous rocks (e.g. limestones or chalk). It is important to know their concentration for the buffering of pH and for issues of water hardness. The buffering capacity of soils, and water derived from soils, is an important concept in water quality. The buffering capacity of a solution is the ability to absorb acid without changing the pH. This is achieved through a high base cation load or high bicarbonate load. This is why soil derived from limestone and chalk has fewer problems coping with acid rain.

Dissolved oxygen

Dissolved oxygen is vital to any aquatic fauna that use gills to breath. Salmonid species of fish require dissolved oxygen contents greater than 5 mg/l, whereas coarse fish (e.g. perch, pike) can survive in levels as low as 2 mg/l. The dissolved oxygen content is also an important factor in the way we taste water. Water saturated in oxygen tastes fresh to human palates; hence drinking water is almost

Table 10.3 Effect of increasing acidity on aquatic ecology

Effect on organisms or process	*pH value*
Mayflies disappear	6.5
Phytoplankton species decline – green filamentous periphyton appears	6
Molluscs disappear	5.5–6.0
Waterfowl breeding declines	5.5
Bacterial decomposition slows/ fungal decomposition appears	5
Salmonid reproduction fails – aluminium toxicity increases	5
Most amphibia disappear	5
Caddis flies, stone flies and Megaloptera (dobsonflies, alderflies, etc.) disappear	4.5–5.0
Beetles, dragonflies and damselflies disappear	4.5
Most adult fish harmed	4.5

Source: Dodds (2002), adapted from Jeffries and Mills (1990)

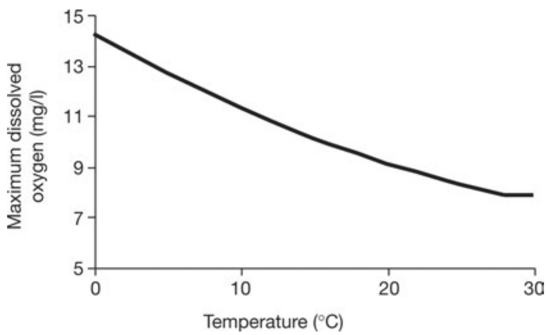


Figure 10.3 Relationship between maximum dissolved oxygen content (i.e. saturation) and temperature.

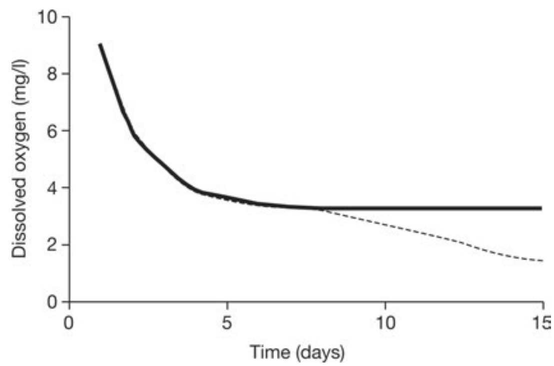


Figure 10.4 Dissolved oxygen curve. The solid line indicates the dissolved oxygen content decreasing due to organic matter. The broken line shows the effect of nitrifying bacteria.

always oxygenated before being sent through a pipe network to consumers.

There are two methods by which dissolved oxygen content is considered: percentage saturation and concentration (mg/l). These two measures are interrelated through temperature, as the dissolved oxygen content of water is highly temperature dependent (see Figure 10.3).

Biochemical oxygen demand

One of the key water-quality parameters is the five-day biochemical oxygen demand test (sometimes referred to as the **biological oxygen demand** test, or BOD_5). This is a measure of the oxygen required by bacteria and other micro-organisms to break down organic matter in a water sample. It is an indirect measure of the amount of organic matter in a water sample, and gives an indication of how much dissolved oxygen could be removed from water as the organic matter decays.

The test is simple to perform and easily replicable. A sample of water needs to be taken, placed in a clean, darkened glass bottle and left to reach 20 °C. Once this has occurred the dissolved oxygen content should be measured (as a concentration). The sample should then be left at 20 °C for 5 days in a darkened environment. After this the dissolved oxygen content should be measured again. The difference between the two dissolved oxygen readings is the BOD_5 value. Over an extended period the dissolved oxygen content of a polluted water sample will look something like that shown in Figure 10.4. In this case the dissolved oxygen content has dropped from 9.0 on day one to 3.6 on day five, giving a BOD_5 value of 5.4 mg/l. After a long period of time (normally more than 5 days) oxygen will start to be consumed by nitrifying bacteria. In this case the bacteria will be consuming oxygen to turn nitrogenous compounds (e.g. ammonium ions) into nitrate. In order to be sure that nitrifying bacteria are not adding to the oxygen demand a suppressant (commonly allyl thiourea or ATU) is added. This ensures that all the oxygen demand is from the decomposition of organic matter. The use of a 5-day period is another safeguard, as, due to the slow growth of nitrifying bacteria, their effect is not noticeable until between 8 and 10 days (Tebbutt 1993). There is an argument to be made saying that it does not matter which bacteria are causing the oxygen demand, the test should be looking at all oxygen demand over a five-day period and therefore there is no need to add ATU. However the standard BOD test uses ATU to suppress the nitrifying bacteria.

In some cases, particularly when dealing with waste water, the oxygen demand will be higher than total saturation. In this case the sample needs to be diluted with distilled water. The maximum dissolved oxygen content at 20 °C is 9.1 mg/l, so any water sample with a BOD_5 value higher than 9 will require dilution. After the diluted test a calculation needs to be performed to find the actual oxygen demand. If you have diluted the sample by half then you need to double your measured BOD_5 value, and so on.

A normal unpolluted stream should have a BOD_5 value of less than 5 mg/l. Untreated sewage is somewhere between 220 and 500 mg/l; while milk has a BOD_5 value of 140,000 mg/l. From these values it is possible to see why a spillage of milk into a stream can have such detrimental effects on the aquatic fauna. The milk is not toxic in its own right, but bacteria consuming the milk will strip the water of any dissolved oxygen and therefore deprive fish of the opportunity to breathe.

There are three reasons why BOD_5 is such a crucial test for water quality:

- Dissolved oxygen is critical to aquatic fauna and the ability to lose dissolved oxygen through organic matter decay is an important measure of stream health.
- It is an indirect measure of the amount of organic matter in the water sample.
- It is the most frequently measured water quality test and has become a standard measure; this means that there are plenty of data to compare readings against.

It also important to realise that BOD is not a direct measure of pollution; rather, it measures the effects of pollution. It also should be borne in mind that there may be other substances present in your water sample that inhibit the natural bacteria (e.g. toxins). In this case the BOD_5 reading may be low despite a high organic load.

Trace organics

Over 600 organic compounds have been detected in river water, mostly from human activity (Tebbutt 1993). Examples include benzene, chlorophenols, pesticides, trihalomethanes and poly-nuclear aromatic hydrocarbons (PAH). These would normally be found in extremely low concentrations but do present significant health risks over the long term. The data for pesticide concentrations (see Table 10.4) in European water resources show that it is a significant problem. This indicates that all water extracted from surface water supplies in Belgium (supplies approximately 30 per cent of the Belgian population) will require pesticide removal before reticulation to customers (Eureau 2001). Although Germany appears to have no pesticide problem, 10 per cent of its surface water resources occasionally have pesticide levels greater than 0.1 μg/l and 90 per cent have pesticides in concentrations less than 0.1 μg/l (but still present) (Eureau 2001).

Some of the trace organic compounds accumulate through the food chain so that humans and other species that eat large aquatic fauna may be at risk. Of particular concern are endocrine disrupting chemicals (EDCs), which have been detected in many rivers. These chemicals, mostly a by-product of industrial processes, attack the endocrine system of humans and other mammals, affecting hormone levels. Some chemicals (e.g. DDT) have the ability to mimic the natural hormone oestrogen. Because oestrogen is part of the reproductive process these chemicals have the potential to affect reproductive

Table 10.4 Percentage of water resources with pesticide concentrations regularly greater than 0.1 μg/l (European Union drinking water standard) for selected European countries

Country	*Surface water (%)*	*Groundwater (%)*
Belgium	100	5.2
Denmark	n/a	8.9
Germany	0.0	0.0
Netherlands	50.0	5.0
UK	77.0	6.0

Source: Data from Eureau (2001)

organs and even DNA. Studies have shown high levels of oestrogen-mimicking compounds in sewage effluent (Montagnani et al. 1996) and that male fish held in cages at sewage effluent discharge sites can develop female sexual organs (Jobling and Sumpter 1993).

Trace organics can be detected using gas chromatography, although this is made difficult by the sheer number of compounds to be detected. They are removed from drinking water supplies using activated carbon filters, or sometimes oxidation by ozone.

Nitrogen compounds

Nitrogen exists in the freshwater environment in four main forms:

- organic nitrogen – proteins, amino acids and urea
- ammonia – either as free ammonia (NH_3) or the ammonium ion (NH_4^+)
- nitrite (NO_2^-)
- nitrate (NO_3^{2-}).

If organic nitrogen compounds enter a river (e.g. in untreated sewage) then an oxidation process called nitrification takes places. An approximation of the process is outlined below:

$$\text{Organic N} + O_2 \rightarrow NH_3/NH_4^+ + O_2 \rightarrow$$
$$NO_2^- + O_2 \rightarrow NO_3^{2-}$$

For this to occur there must be nitrifying bacteria and oxygen present. This is one of the main processes operating in a sewage treatment works (see pp. 225–227) – the breakdown of organic nitrogenous compounds into a stable and relatively harmless nitrate. There are two problems with this process occurring in the natural river environment. First, there is the oxygen demand created by the nitrification process. Second, the intermediate ammonia stage is highly toxic, even in very low concentrations. Under extremely low dissolved oxygen concentrations (less than 1 mg/l) the nitrification process can be reversed, at least in the first stage. In this case nitrates will turn into nitrite and oxygen will be released. Unfortunately, this is not a ready means for re-oxygenating a river as by the time the dissolved oxygen level has dropped to 1 mg/l the fish population will have died or moved elsewhere.

The levels of nitrate in a water sample can be expressed in two different ways: absolute nitrate concentration, or the amount of nitrogen held as nitrate (normally denoted as NO_3–N). The two are related by a constant value of approximately 4.4. As an example, the World Health Organisation recommended that the drinking water standard for nitrate in drinking water be 45 mg/l. This can also be expressed as 10 mg/l NO_3–N.

As indicated above, one source of nitrate is from treated sewage. A second source is from agricultural fertilisers. Farmers apply nitrate fertilisers to enhance plant growth, particularly during the spring. Plants require nitrogen to produce green leaves, and nitrates are the easiest form to apply as a fertiliser. This is because nitrates are extremely soluble and can easily be taken up by the plant through its root system. Unfortunately this high solubility makes them liable to be flushed through the soil water system and into rivers. To make matters worse a popular fertiliser is ammonium nitrate – $(NH_4)_2NO_3$. This has the added advantage for the farmer of three nitrogen atoms per molecule. It has the disadvantage for the freshwater environment of extremely high solubility and providing ammonium ions in addition to nitrate. The application of nitrate fertilisers is most common in areas of intensive agricultural production such as arable and intensive livestock farming.

Another source of nitrates in river systems is from animal wastes, particularly in dairy farming where slurry is applied to fields. This is organic nitrogen (frequently with high urea content from urine) which will break down to form nitrates. This is part of the nitrification process described earlier.

A fourth source of nitrates in river systems is from plants that capture nitrogen gas from the air. This is not strictly true, as it is actually bacteria such as Rhizobium, attached to a plant's root, that capture the gaseous nitrogen and turn it into water-soluble forms for the plants to use. Not all plants have this ability; in agriculture it is the legumes, such as clovers, lucerne (or alfalfa), peas and soy beans, that can gain nitrogen in this way. Once the nitrogen is in a soluble form it can leach through to a river system in the same way that fertilisers do. Over a summer period the nitrogen levels in a soil build up and then are washed out when autumn and winter rains arrive. This effect is exacerbated by ploughing in the autumn, which releases large amounts of soil-bound nitrogen.

There is one other source of nitrates in rivers: atmospheric pollution. Nitrogen gas (the largest constituent of the atmosphere) will combine with oxygen whenever there is enough energy for it to do so. This energy is readily supplied by combustion engines (cars, trucks, industry, etc.) producing various forms of nitrogen oxide gases (often referred to as NO_x gases). These gases are soluble to water in the atmosphere and form nitrites and nitrates in rainwater. This is not a well-studied area and it is difficult to quantify how much nitrogen reaches rivers from this source (see p. 47).

The different sources of nitrate in a river add together to give a cycle of levels to be expected in a year. Figure 10.5 shows this cycle over a 3-year period on the river Lea, south-east England. The low points of nitrate levels correspond to the end of a summer period, with distinct peaks being visible over the autumn to spring period, particularly in the spring. The Lea is a river that has intensive arable agriculture in its upper reaches, but also a significant input from sewage effluent. At times during the summer months the Lea can consist of completely recycled water, and the water may have been through more than one sewage works. This gives a background nitrate level, but it is perhaps surprising that the summer levels of nitrate are not higher, compared to the winter period. Partly this can be attributed to the growth of aquatic plants in the summer, which remove nitrate from the water. The peaks over the autumn–spring period are as a result of agricultural practices discussed above. The example given here is specific to the south-east of England; in different parts of the world the cycles will differ in timing and extent.

Nitrates are relatively inert and do not create a major health concern. An exception to this is methaemoglobinaemina ('blue baby syndrome'). Newborn babies do not have the bacteria in their stomach to deal with nitrates in the same manner as older children and adults. In the reducing surroundings of the stomach the nitrate is transformed into nitrite that then attaches itself to the haemoglobin molecule in red blood cells, preferentially replacing oxygen. This leads to a reduction in oxygen supply around the body, hence the name 'blue baby syndrome'. In reality methaemoglobinaemina is extremely rare, possibly coming from nitrate-polluted well supplies but not mains-supplied drinking water. The drinking water limit for the European Union is 50 mg/l of nitrate (44 mg/l in the USA). In rivers it is rare to have nitrate values as high as this. In a study of streams draining intensively dairy-farmed land in the North Island of New Zealand, Rodda et al. (1999) report maximum nitrate levels of 26.4 mg/l. These are reported as being 'very high by New Zealand standards' (Rodda et al. 1999: 77). In Figure 10.5 the peak nitrate level for the river Lea in England is 21 mg/l, with the norm being somewhere between 5 and 10 mg/l.

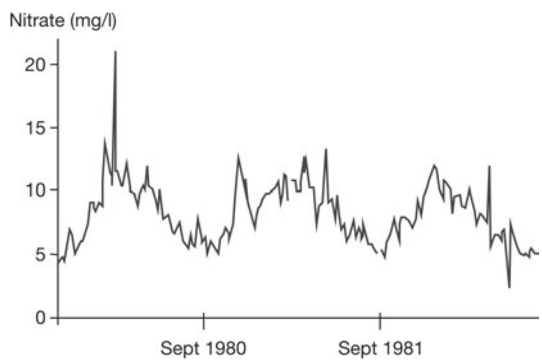


Figure 10.5 Nitrate levels in the river Lea, England. Three years of records are shown: from September 1979 until September 1982.

Source: Data from the Environment Agency

The biggest concern with nitrates in a river system is **eutrophication**. In exactly the same way that the nitrogen enhances the growth of land-based plants, it will also boost the growth of aquatic plants, including algae. This creates a problem of over-production of plant matter in river systems. This is discussed in more detail on pp. 224–225.

Phosphates

Phosphorus can be found in three different forms: orthophosphate, polyphosphate (both normally dissolved) and organic phosphate (bound to organic particles). The ratio of different forms of phosphorus in a water sample is highly pH dependent (Chapman 1996). Like nitrogen, the availability of phosphorus is a limiting factor in plant growth. The most common form of application for plants is as phosphate. The major difference from nitrates is that phosphate is not nearly as soluble. Consequently phosphate is normally applied as a solid fertiliser, and less frequently than nitrate. In river systems the main source of dissolved phosphate is from detergents and soaps that come through sewage treatment works. Sewage treatment works remove very little of the phosphate from detergents present in waste water, except where specific phosphate-stripping units are used. The largest amount of phosphate in river systems is normally attached to particles of sediment. Rodda et al. (1999) report maximum dissolved reactive phosphorus levels of 0.2 mg/l but total phosphorus levels of 1.6 mg/l. This is for intensive dairy production, where the majority of phosphate is from agricultural fertilisers.

Phosphates are a major contributor to eutrophication problems. The fact that they are bound to sediments means that they often stay in a river system for a long period of time. Improvements in water quality for a river can often be delayed substantially by the steady release of phosphate from sediments on the river bed.

Chlorine

Chlorine is not normally found in river water. It is used as a disinfectant in the supply of drinking water. It is used because it is toxic to bacteria and relatively short lived. More common to find in river water samples is the chloride ion. This may be an indicator of sewage pollution as there is a high chloride content in urine. Chloride ions give the brackish taste of sea water, the threshold for taste being around 300 mg/l. The European Commission limit for drinking water is 200 mg/l.

Heavy metals

'Heavy metals' is the term applied to metals with an atomic weight greater than 6. They are generally only found in very low levels dissolved in fresh water, but may be found in bed load sediments. In acidic waters metals can be dissolved (i.e. found in ionic form). They are often toxic in concentrations above trace levels. The toxicity, in decreasing order, is mercury, cadmium, copper, zinc, nickel, lead, chromium, aluminium and cobalt (Gray 1999). In the aquatic environment, copper and zinc are the most frequent causes of toxicity. A major source of zinc is derived from galvanised steel, particularly in wire fencing and roofs (Alloway and Ayres 1997).

Accumulation of lead in sediments has been a problem for aquatic wildlife. Since the banning of leaded petrol, the major source has been through the use of lead shot and fishing sinkers. Lead shot has been banned in favour of steel shot in many countries (e.g. USA, UK, New Zealand, Australia) due to these problems (Dodds 2002).

The sources of heavy metals in the aquatic environment are almost always industrial or surface runoff from roads. Sewage sludge (the product of sedimentation at a sewage treatment works) is frequently heavy metal-rich, derived from industry discharging waste into the sewerage system. When untreated sewage is discharged into a river, heavy metals can be found in the sediments. Where there is a combined sewage and stormwater drainage system for an urban area, untreated sewage can be discharged during a storm event when the sewage treatment works cannot cope with the extra storm water. Runoff from roads (through a stormwater system) frequently shows high levels of copper from vehicle brake pads. When washed into a river system, particularly in summer storms, the copper levels can be extremely high and cause toxicity problems to aquatic fauna.

WATER-QUALITY MEASUREMENT

The techniques used for water-quality analysis vary considerably depending on equipment available and the accuracy of measurement required. For the highest accuracy of measurement, water samples should be taken back to a laboratory, but this is not always feasible. There are methods that can be carried out in the field to gain a rapid assessment of water quality. Both field and laboratory techniques are discussed on the following pages. Before discussing the measurement techniques it is important to consider how to sample for water quality.

Sampling methodology

It is difficult to be specific on how frequently a water sample should be taken, or how many samples represent a given stretch of water. The best way of finding this out is to take as many measurements as possible in a trial run. Then statistical analyses can be carried out to see how much difference it would have made to have had fewer measurements. By working backwards from a large data set it is possible to deduce how few measurements can be taken while still maintaining some accuracy of overall assessment. An example of this type of approach, when used for the reduction in a hydrometry (water quantity) network, is in Pearson (1998). The main concern is that there are enough measurements to capture the temporal variability present and that the sample site is adequately representative of your river stretch.

One important consideration that needs to be understood is that the sample of water taken at a particular site is representative of all the catchment above it, not just the land use immediately adjacent. Adjacent land use may have some influence on the water quality of a sample, but this will be in addition to any affect from land uses further upstream which may be more significant.

Gravimetric methods

Gravimetric analysis depends on the weighing of solids obtained from a sample by evaporation, filtration or precipitation (or a combination of these three). This requires an extremely accurate weighing balance and a drying oven, hence it is a laboratory technique rather than a field one. An example of gravimetric analysis is the standard method for measuring total dissolved solids (TDS). This is to filter a known volume of filter paper. The sample of water is then dried at 105 °C and the weight of residue left is the TDS.

Other examples of gravimetric analysis are total suspended solids and sulphates (causing a precipitate and then weighing it).

Volumetric methods

Volumetric analysis is using titration techniques to find concentrations of designated substances. It is dependent on measuring the volume of a liquid reagent (of known concentration) that causes a visible chemical reaction. This is another laboratory technique as it requires accurate measurements of volume using pipettes and burettes. Examples of this technique are chloride and dissolved oxygen (using the Winkler method).

Colorimetry

Colorimetric analysis depends on a reagent causing a colour to be formed when reacting with the particular ion you are interested in measuring. The strength of colour produced is assumed to be proportional to the concentration of the ion being measured (Beer's law). The strength of colour can then be assessed using one of four techniques: comparison tubes, colour discs, colorimeter or spectrophotometer.

Comparison tubes are prepared by using standard solutions of the ion under investigation which the reagent is added to. By having a range of standard solutions the strength of colour can be compared (by eye) to find the concentration of the water sample. The standard solutions will fade with time and need remaking, hence this is a time-consuming method.

Colour discs use the same principle as comparison tubes, except in this case the standards are in the form of coloured glass or plastic filters. The coloured sample is visually compared to the coloured disc to find the corresponding concentration. It is possible to buy colour disc kits that come with small packets of reagent powder for assessment of a particular ion. This method is extremely convenient for rapid field assessment, but is subjective and prone to inaccuracy.

A colorimeter (sometimes called an absorptiometer) takes the subjective element out of the assessment. It is similar to a turbidity meter in that a beam of light is shone through the reagent in a test tube. The amount of light emerging from the other side is detected by a photo-electric cell. The darker the solution (caused by a high concentration of reactive ion) the less light emerges. This reading can then be compared against calibrations done for standard solutions.

A spectrophotometer is the most sophisticated form of colorimetric assessment. In this case instead of a beam of white light being shone through the sample (as for the colorimeter) a specific wavelength of light is chosen. The wavelength chosen will depend on the colour generated by the reagent and is specified by the reagent's manufacturer.

There are a range of spectrophotometers available to perform rapid analysis of water quality in either a laboratory or field situation. Many ions of interest in water-quality analysis can be assessed using colorimetric analysis. These include nitrate, nitrite, ammonia and phosphate.

Ion-selective electrodes

In a similar vein to pH meters, ion-selective electrodes detect particular ions in solution and measure the electrical potential produced between two reactive substances. The tip of the electrode in the instrument has to be coated with a substance that reacts with the selected ion. With time the reactive ability of the electrode will decrease and need to be replaced. Although convenient for field usage and accurate, the constant need for replacing electrodes makes these an expensive item to maintain. There are ion-selective electrodes available to measure dissolved oxygen, ammonium, nitrate, calcium, chloride and others.

Spectral techniques

When ions are energised by passing electricity through them, or in a flame, they produce distinctive colours. For instance, sodium produces a distinctive yellow colour, as evidenced by sodium lamps used in some cars and street lamps. Using spectral analysis techniques, the light intensity of particular ions in a flame are measured and compared to the light intensity from known standard solutions. The most common form of this analysis is atomic absorption spectrophotometry, a laboratory technique which is mostly used for metallic ions.

PROXY MEASURES OF WATER QUALITY

Any measurement of water quality using individual parameters is vulnerable to the accusation that it represents one particular point of time but not the overall water quality. It is often more sensible to try and assess water quality through indirect measurement of something else that we know is influenced by water quality. Two such proxy measures of water quality are provided by biological indicators and analysis of sediments in the river.

Biological indicators

Aquatic fauna normally remain within a stretch of water and have to try and tolerate whatever water pollution may be present. Consequently the health of aquatic fauna gives a very good indication of the water quality through a reasonable period of time. There are two different ways that this can be done: catching fauna and assessing their health; or looking for the presence and absence of key indicator species.

Fish surveys are a common method used for assessing the overall water quality in a river. It is an expensive field technique as it requires substantial human resources: people to wade through the water with electric stun guns and then weigh and measure stunned fish. When this is done regularly it gives very good background information on the overall water quality of a river.

More common are biological surveys using indicator species, particularly of macro-invertebrates. Kick sampling uses this technique. A bottom-based net is kicked into sediment to catch any bottom-based macro-invertebrates, which are then counted and identified. There are numerous methods that can be used to collate this species information. In Britain the BMWP (Biological Monitoring Working Party) score is commonly used and provides good results. Species are given a score ranging from 1 to 10, with 10 representing species that are extremely intolerant to pollution. The presence of any species is scored (it is purely presence/absence, not the total number) and the total for the kick sample calculated. The BMWP score has a maximum of 250. Other indicator species scores include the Chandler index and the ASPT (Average Score Per Taxon). Details of these can be found in a more detailed water-quality assessment text such as Chapman (1996).

Another example of an indicator species used for water-quality testing is *Escherichia coli* (*E. coli*). These are used to indicate the presence or absence of faecal contamination in water. *E. coli* is a bacteria present in the intestines of all mammals and excreted in large numbers in faeces. Although one particular strain (*E. coli*$_{157}$) has toxic side effects, the vast majority of *E. coli* are harmless to humans. Their presence in a water sample is indicative of faecal pollution, which may be dangerous because of other pathogens carried in the contaminated water. They are used as an indicator species because they are easy to detect, while viruses and other pathogens are extremely difficult to measure. Coliform bacteria (i.e. bacteria of the intestine) are detected by their ability to ferment lactose, producing acid and gas (Tebbutt 1993). There are specific tests to grow *E. coli* in a lactose medium, which allow the tester to derive the most probable number per 100 ml (MPN/100 ml).

Sediments

The water in a channel is not the only part of a river that may be affected by water pollution. There are many substances that can build up in the sediments at the bottom of a river and provide a record of pollution. There are two big advantages to this method for investigating water quality: the sediments will reflect both instantaneous large pollution events and long, slow contamination at low levels; and if the river is particularly calm in a certain location the sediment provides a record of pollution with time (i.e. depth equals time). Not all water pollutants will stay in sediments, but some are particularly well suited to study in this manner (e.g. heavy metals and phosphorus).

The interpretation of results is made difficult by the mobility of some pollutants within sediments. Some metals will bind very strongly to clay particles in the sediments (e.g. lead and copper), and you can be fairly certain that their position is indicative of where they were deposited. Others will readily disassociate from the particles and move around in the interstitial water (e.g. zinc and cadmium) (Alloway and Ayres 1997). In this case you cannot be sure that a particularly high reading at one depth is from deposition at any particular time.

MODELLING WATER QUALITY

The numerical modelling of water quality is frequently required, particularly to investigate the effects of particular water-quality scenarios. The type of problems investigated by modelling are: the impact of certain levels of waste discharge on a river (particularly under low flow levels); recovery of a water body after a pollution event; the role of backwaters for concentration of pollutants in a river; and many more. The simplest water-quality models look at the concentration of a certain pollutant in a river given knowledge about flow conditions and decay rates of the pollutant. The degradation of a pollutant with time can be simulated as a simple exponential decay rate equation. A simple mass balance approach can then be used to calculate the amount of pollutant left in the river after a given period of time (James 1993). More complex models build on this approach and incorporate ideas of diffusion, critical loads of pollutants and chemical reaction between pollutants in a river system. If the problem being researched is to track pollutants down a river then it is necessary to incorporate two- or three-dimensional representation of flow hydraulics. There are numerous water-quality models available in the research literature, as well as those used by consultants and water managers.

EUTROPHICATION

'Eutrophication' is the term used to describe the addition of nutrients to an aquatic ecosystem that leads to an increase in net primary productivity. The term comes from limnology (the study of freshwater bodies, e.g. lakes and ponds) and is part of an overall classification system for the nutrition, or trophic, level of a freshwater body. The general classification moves from oligotrophic (literally 'few nutrients'), to eutrophic ('good nutrition') and ends with hypertrophic ('excess nutrients'). In limnology this classification is viewed as part of a natural progression for bodies of water as they fill up with sediment and plant matter. Eutrophication is a natural process (as part of the nitrogen and phosphorus cycles), but it is the addition of extra nutrients from anthropogenic activity that attracts the main concern in hydrology. In order to distinguish between natural and human-induced processes the term 'cultural eutrophication' is sometimes used to identify the latter.

The major nutrients that restrict the extent of a plant's growth are potassium (K), nitrogen (N) and phosphorus (P). If you buy common fertiliser for a garden you will normally see the N:P:K ratio expressed to indicate the strength of the fertiliser. For both aquatic and terrestrial plants nitrogen is required for the production of chlorophyll and green leaves, while potassium and phosphorus are needed for root and stem growth. In the presence of abundant nitrogen and phosphorus (common water pollutants, see pp. 218–220), aquatic plant growth, including algae, will increase dramatically. This can be seen as positive as it is one way of removing the nitrate and phosphate from the water, but overall it has a negative impact on the river system. The main negative effect is a depletion of dissolved oxygen caused by bacteria decomposing dead vegetative matter in the river. In temperate regions this is a particular problem in the autumn when the aquatic vegetation naturally dies back. In tropical regions it is a continual problem. A second negative effect is from algal blooms. In 1989 there was an explosion in numbers of cyanobacteria in Rutland Water, a reservoir supplying drinking water in central England (Howard 1994). (NB These are also called blue-green algae, despite being a phylum of bacteria.) Cyanobacteria produce toxins as waste products of respiration that can severely affect water quality. In the 1989 outbreak several dogs and sheep that drank water from Rutland Water were poisoned, although no humans were affected (Howard, 1994). In an effort to eliminate future problems, the nutrient-rich source water for Rutland Water is supplemented with water from purer river water pumped from further afield.

Eutrophication of water can occur at what appear to be very low nutrient levels. As an example the drinking water standard for nitrate-nitrogen is around 12 mg/l (depending on country) but concentrations as low as 2–3 mg/l can cause eutrophication problems in water bodies.

Table 10.5 shows some of the indicators used in a quantitative example of defined trophic levels developed for the Organisation for Economic Cooperation and Development (OECD). The

Table 10.5 OECD classification of lakes and reservoirs for temperate climates

Trophic level	*Average total P (mg/l)*	*Dissolved oxygen (% saturation)*	*Max. chlorophyll (mg/l) (at depth)*
Ultra-oligotrophic	0.004	>90	0.0025
Oligotrophic	0.01	>80	0.008
Mesotrophic	0.01–0.035	40–89	0.008–0.025
Eutrophic	0.035–0.1	0–40	0.025–0.075
Hypertrophic	>0.1	0–10	>0.075

Source: Adapted from Meybeck et al. (1989)

chlorophyll is an indicator of algal growth in the water, while phosphorus and dissolved oxygen are more traditional water quality measures. The dissolved oxygen is taken from the bottom of the lake because this is where vegetative decomposition is taking place. The dissolved oxygen level near to the surface will vary more because of the proximity to the water/air interface and the oxygen produced in photosynthesis by aquatic plants. It is worth noting that heavily eutrophied water samples will sometimes have a dissolved oxygen greater than 100 per cent. This is due to the oxygen being produced by algae which can supersaturate the water.

CONTROLLING WATER QUALITY

Waste water treatment

The treatment of waste water is a relatively simple process that mimics natural processes in a controlled, unnatural environment. The treatment processes used for industrial waste water are dependent on the type of waste being produced. In this section the processes described are those generally found in sewage treatment rather than in specialised industrial waste water treatment.

There are two major objectives for successful sewage treatment: to control the spread of disease from waste products and to break down the organic waste products into relatively harmless metabolites (i.e. by-products of metabolism by bacteria, etc.). The first objective is achieved by isolating the waste away from animal hosts so that viruses and other pathogens die. The second objective is particularly important for the protection of where the treated effluent ends up – frequently a river environment.

In Britain the first attempt to give guidelines for standards of sewage effluent discharge were provided by the Royal Commission on Sewage Disposal which sat between 1898 and 1915. The guidelines are based on two water-quality parameters described earlier in this chapter: suspended solids and biochemical oxygen demand (BOD). The Royal Commission set the so-called 30:20 standard which is still applicable today (i.e. 30 mg/l of suspended solids and 20 mg/l of BOD). The standard was based on a dilution ratio of 8:1 with river water. Where river flow is greater than eight times the amount of sewage effluent discharge the effluent should have a TSS of less than 30 mg/l and a BOD of less than 20 mg/l. There was also the recommendation that if the river is used for drinking water extraction further downstream the standard should be tightened to 10:10. This was used as a recommendation until the 1970s when a system of legal consents to discharge was introduced (see p. 228).

The processes operating at a waste water treatment works are very simple. They are summarised below and in Figure 10.6 (NB not every sewage treatment works will have all of these processes present).

1 Primary treatment: screening and initial settlement.
2 Secondary treatment: encouraging the biological breakdown of waste and settling out of remaining solids. This can take place either in trickle bed filters or activated sludge tanks. The main requirement is plenty of oxygen to allow micro-organisms to break down the concentrated effluent.

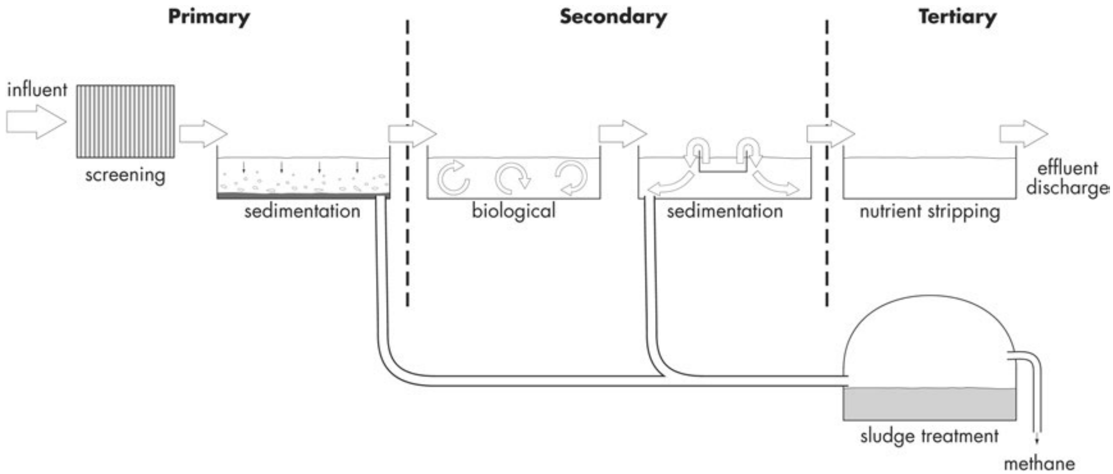


Figure 10.6 Schematic representation of waste water treatment from primary through to tertiary treatment, and discharge of the liquid effluent into a river, lake or the sea.

3 Tertiary treatment: biodigestion of sludge (from earlier settling treatment); extra treatment of discharging effluent to meet water-quality standards (e.g. phosphate stripping, nitrate reduction).

Raw sewage entering a sewage treatment works is approximately 99.9 per cent water (Gray 1999). This is derived from water used in washing and toilet flushing, and also from storm runoff in an urban environment where there is a combined sewage/stormwater drainage scheme. Of the solids involved, the majority are organic and about half are dissolved in the water (TDS). Of the organic compounds the breakdown is approximately 65 per cent nitrogenous (proteins and urea), 25 per cent carbohydrates (sugars, starches, cellulose) and 10 per cent fats (cooking oils, grease, soaps) (Gray 1999). Typical values for TSS and BOD at different stages of sewage treatment are provided in Table 10.6.

Table 10.6 Changes in suspended solids and biochemical oxygen demand through sewage treatment. These are typical values which will vary considerably between treatment works

Stage of treatment	*Suspended solids (mg/l)*	*BOD (mg/l)*
Raw sewage	400	300
After primary treatment	150	200
After biological treatment	300	20
Effluent discharged to river	30	20

In tertiary treatment an effort is sometimes (but not always) made to reduce the level of nitrate and phosphorus in the discharged waste. In some cases this is achieved through final settling ponds where the growth of aquatic flora is encouraged and the nutrients are taken up by the plants before discharge into a stream. Of particular use are reeds which do not die back during the winter (in temperate regions). This is a re-creation of natural wetlands that have been shown to be extremely efficient removers of both nitrogen and phosphorus from streams (e.g. Russel and Maltby 1995). Other methods of phosphate removal are to add a lime or metallic salt coagulant that causes a chemical reaction with the dissolved phosphorus so that an insoluble form of phosphate settles out. This is particularly useful where the receiving water for the final effluent has problems with eutrophication. The average phosphorus concentration in raw sewage is 5–20 mg/l, of which only 1–2 mg/l is removed in biological treatment.

In some cases, particularly in the USA, chlorination of the discharging effluent can take place. Chlorine is used as a disinfectant to kill any pathogens left after sewage treatment. This is a noble aim but creates its own difficulties. The chlorine can attach to organic matter left in the effluent and create far worse substances such as polychlorinated biphenyl (PCB) compounds. Another safer form of disinfection is to use ultraviolet light, although this can be expensive to install and maintain.

Source control

The best way of controlling any pollution is to try and prevent it happening in the first place. In order to achieve this differentiation has to be made between point source and diffuse pollutants (see p. 211). When control over the source of pollutants is achieved, dramatic improvements in river-water quality can be achieved. An example of this is shown in the Case Study of the Nashua River in Massachusetts, USA.

Case study

CONTROLLING WATER QUALITY OF THE NASHUA RIVER

The Nashua river is an aquatic ecosystem that has undergone remarkable change in the last 100 years. It drains an area of approximately 1,400 km^2 in the state of Massachusetts, USA, and is a tributary of the much larger Merrimack river which eventually flows into the sea in Boston Harbor (see Figure 10.7). The land use of the Nashua catchment is predominantly forest and agricultural, with a series of towns along the river. It is the industry associated with these towns that has brought about the changes in the Nashua, predominantly through the twentieth century. The latter-day changes are well illustrated by the two photographs at the same stretch of the Nashua, in 1965 and 1995 (see Figures 10.8 and 10.9).

Prior to European colonisation of North America the Nashua valley was home to the Nashaway tribe, and the Nashua river could be considered to be in a pristine condition. With the arrival of European settlers to New England the area was used for agriculture and the saw milling of the extensive forests. The Industrial Revolution of the nineteenth century brought manufacturing to the area and mills sprang up along the river. By the middle of the twentieth century the small towns along the Nashua (Gardner, Fitchburg, Leominster and Nashua) were home to paper, textile

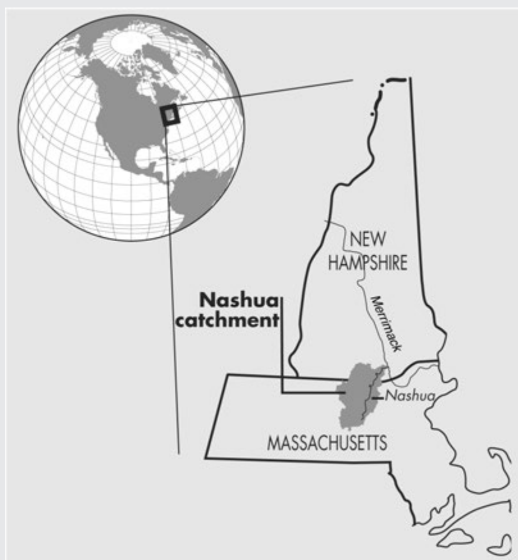


Figure 10.7 Location of the Nashua catchment in north-east USA.

and shoe factories, many of which were extracting water from the river and then discharging untreated waste back into the river. The photograph of the Nashua in 1965 (Figure 10.8) is indicative of the pollution problems experienced in the river; in this case dye from a local paper factory has turned the river red. Under the US water-quality classification scheme the river was classified as U: unfit to receive further sewage.

In 1965 the Nashua River Clean-Up Committee was set up to try to instigate a plan of restoring the water quality in the river. This committee later became the Nashua River Watershed Association (NRWA) which still works today to improve water-quality standards in the area. Between 1972 and 1991, 11 waste water treatment plants were constructed or upgraded to treat waste from domestic, and to a lesser extent from industrial, sources in the catchment. These were built using grants from the state and federal government as part of a strategy to improve the river from U to B status (fit for fishing and swimming). Through this control of point source pollution, the river-water quality has improved dramatically as can be seen in the second photograph of the river (Figure 10.9). The river has attained B status and is an important recreational asset for the region. It has not returned to a pristine state, though, and is unlikely to while there is still a significant urban population in the catchment. There are problems with combined sewage and stormwater drainage systems discharging untreated waste into the river during large storms, and also diffuse pollution sources – particularly in the urban environment. However, during the latter half of the twentieth century the Nashua river has had its water quality transformed from an abiotic sewer into a clean river capable of maintaining a healthy salmonid fish population. This has largely been achieved through the control of point pollution sources.

The authors gratefully acknowledge the Nashua River Watershed Association for supplying much of this information and Figures 10.8 and 10.9. For more information on the NRWA visit:www.nashuariverwatershed.org.

Figure 10.8 The Nashua river during 1965, prior to water pollution remediation measures being taken.

Source: Photo courtesy of the Nashua River Watershed Association

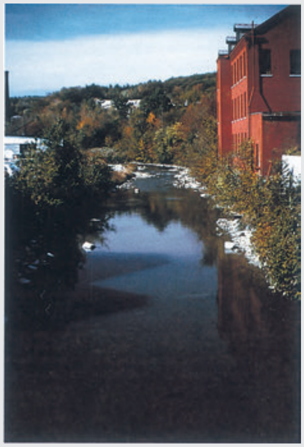


Figure 10.9 The Nashua river during the 1990s, after remediation measures had been taken.

Source: Photo courtesy of the Nashua River Watershed Association

Controlling point source pollutants

The control of point source pollutants cannot always be achieved by removing that point source. It is part of water resource management to recognise that there may be valid reasons for the disposal of waste in a river; effective management ensures that waste disposal creates no harmful side effects. In the United Kingdom the control of point source pollution is through discharge consents. These provide a legal limit for worst-case scenarios – for example, at individual sewage treatment works they are usually set with respect to TSS, BOD and ammonia (sometimes heavy metals are included), and calculated to allow for low flow levels in the receiving stream (see the technique box below for calculating discharge consents). There is also an obligation to comply with the European Union Urban Waste Water directive.

Technique: Calculating discharge consents

In England, the setting of discharge consents for point source pollution control is carried out by the Environment Agency. A discharge consent gives a company the right to dispose of a certain amount of liquid waste into a river system so long as the pollution levels within the discharge are below certain levels. To calculate what those critical levels are, a series of computer programs are used. These computer programs are in the public domain and can be obtained from the Environment Agency. They use very simple principles that are described here.

The main part of the discharge consent calculation concentrates on a simple mass balance equation:

$$C_D = \frac{Q_U C_U + Q_E C_E}{Q_B + Q_E} \tag{10.3}$$

Where Q refers to the amount of flow (m^3/s) and C the concentration of pollutant. For the subscripts: D is for downstream; U is for upstream (i.e. the background); and E is for the effluent.

With this mass balance equation, the downstream concentration can be calculated with varying flows and levels of effluent concentrations. This variation in flow and concentration is achieved through a computer program running a Monte Carlo simulation.

In this case the Monte Carlo simulation involves a random series of values for Q_U, Q_E, C_U and C_E drawn from an assumed distribution for each variable. It is assumed that the distributions are log-normal in shape (see Figure 10.10) and therefore using the data in Table 10.7 the actual distribution for each variable is simulated. Once the distribution for each variable is known then a random variable is chosen from that distribution. In the case of a log-normal distribution this means that it is most likely to be close to the mean value but more likely to be above than below the mean (see Figure 10.10). In a Monte Carlo simulation the value of C_D is calculated many times (often set to 1,000) so that a distribution for C_D can be drawn. The consent to discharge figure is taken from the distribution of C_D, usually looking at the 90 or 95 percentile

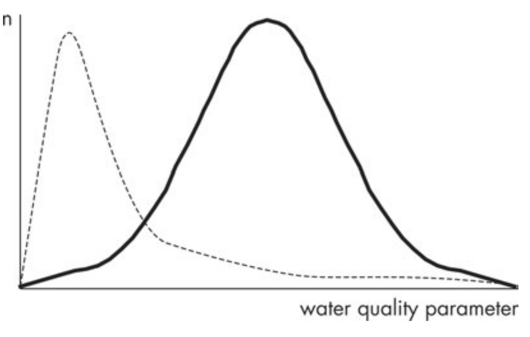


Figure 10.10 A log-normal distribution (broken line) compared to a normal distribution (solid line).

Table 10.7 Parameters required to run a Monte Carlo simulation to assess a discharge consent

Variable	*Required data*
River flow (Q_U)	Mean daily flow and Q_{95}
Upstream river quality (C_U)	Mean value and standard deviation
Effluent flow (Q_E)	Mean value and standard deviation
Effluent quality (C_E)	Mean and standard deviation

values, i.e. the target will be achieved 90 or 95 per cent of the time.

The values that are required for calculating a consent to discharge (see Table 10.7) are derived from normal hydrological data. River flow data can be derived from a flow duration curve (see Chapter 9). The water quality information requires at least 3 to 4 years of regular measurements. The values to describe Q_E and C_E will either be known or are to be varied in the simulation in order to derive a consent to discharge value.

In short, the person calculating the discharge consent inserts values from Table 10.7 into the Monte Carlo simulation. This will then produce the 95 percentile value of C_D. If that value is too high (i.e. too much pollution) then the simulation is run again using lower values for C_E until a reasonable value is derived. Once the reasonable value has been reached then the 95 percentile value of C_E is taken as the discharge consent. The definition of a 'reasonable value' will be dependent on the designated use of the river. Rivers with high-class fisheries and those with abstraction for potable supply have much higher standards than for other uses.

The approach described here can be used to calculate a consent to discharge for such water quality parameters as BOD and TSS. When a calculation is being carried out for ammonia then more data are required to describe the water quality in the receiving river. Parameters such as pH, temperature, alkalinity, TDS and dissolved oxygen (all described with mean and standard deviation values) are required so that chemical reaction rates within the river can be calculated.

A scheme such as discharge consents provides a legal framework for the control of point source pollutants but the actual control comes about through implementing improved waste water treatment.

Controlling diffuse source pollutants

The control of diffuse source water pollution is much harder to achieve. In an urban environment this can be achieved through the collection of stormwater drainage into artificial wetlands where natural processes can lessen the impact of the pollutants on the draining stream. Of particular concern is runoff derived from road surfaces where many pollutants are present as waste products from vehicles. Hamilton and Harrison (1991) suggest that although roads only make up 5–8 per cent of an urban catchment area they can contribute up to 50 per cent of the TSS, 50 per cent of the total hydrocarbons and 75 per cent of the total heavy metals input into a stream. The highest pollutant loading comes during long, dry periods which may be broken by flushes of high rainfall (e.g. summer months in temperate regions). In this case the majority of pollutants reach the stream in the first flush of runoff. If this runoff can be captured and held then the impact of these diffuse pollutants is lessened. This is common practice for motorway runoff where it drains into a holding pond before moving into a nearby water course.

Another management tool for control of diffuse pollutants is to place restrictions on land management practices. An example of this is in areas of England that have been designated either a Nitrate Vulnerable Zone (NVZ) or a Nitrate Sensitive Area (NSA), predominantly through fears of nitrate contamination in aquifers. In NSAs the agricultural practices of muck spreading and fertilising with nitrates are heavily restricted. This type of control relies on tight implementation of land use planning – something that is not found uniformly between countries, or even within countries.

Examples of controlling water pollution

Biggs (1989) presents data showing the recovery of a river in New Zealand following effective treatment of a point source pollution problem. The pollution was due to discharge of untreated effluent from an abattoir, directly into a nearby branch of the Waimakariri river. Water quality was monitored upstream and downstream of the discharge point before and after a staged improvement in wastewater treatment at the abattoir. The results shown in Figure 10.11 use an autotrophic index, a ratio of the periphyton mass to chlorophyll-a. This is a measure of the proportions of heterotrophic (require organic carbon to survive) to autotrophic (produce organic compounds from simple molecules) organisms. The time series of data upstream and downstream from the abattoir (Figure 10.11) can be split between the pretreatment period (Sep–Oct 1985), the recovery period (May–Aug 1986) and the recovered period (after August 1986) (Biggs 1989). (NB the vertical axis is on a logarithmic scale so differences appear smaller.) A remarkable point about this study is how quickly the river appears to have recovered (approximately 5 months) following treatment of the point source pollution. This is a reflection of the low residence time of the pollutants within the river system and the effective flushing out of the pollutants by the river.

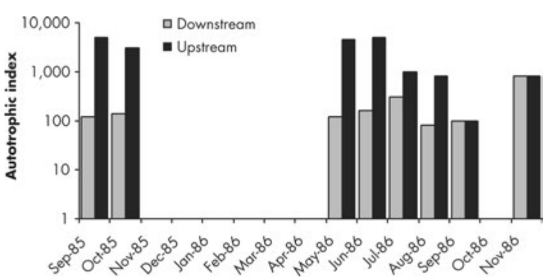


Figure 10.11 Recovery in water quality after improved waste water treatment at an abattoir. The waste water treatment was implemented with progressive reductions in effluent discharged into the river from May 1986. See text for explanation of vertical axis.

Source: Redrawn from Biggs (1989)

Dodds (2002) presents two case studies on lake eutrophication with varying degrees of success. The first is for Lake Washington on the eastern border of Seattle, USA. For Lake Washington the diversion of treated sewage away from the lake (achieved in 1963) was enough to halt the decline in water quality and return the lake to an oligotrophic state. For Lake Trummen in Sweden the stopping of sewage input into the lake was not enough to improve water quality since high levels of phosphorous remained in the lake sediment continuing the eutrophication problem. In this case a dramatic rise in water quality was achieved by dredging the lake sediments (and selling the dredged sediments as nutrient topsoil) so that the lake was able to be returned to recreational usage (Dodds 2002).

In New Zealand there is an ongoing study to improve the water quality in Lake Rotorua in the Central North Island. This is a lake of tremendous importance for tourism and of great cultural importance to the local Maori people. Initially it was thought that the water quality problem could be solved through treating the point source pollution at a sewage treatment plant which received a significant upgrade in 1990. Although this caused a temporary decrease in nutrient loading to the lake the water quality has continued to decline, largely due to agricultural intensification in the lake catchment area. Nitrate-nitrogen levels in the streams feeding into the lake are in the order of 1–2 mg/l but have increased significantly over the past 30 years (White et al. 2007). A major concern is that the groundwater levels of nitrate-nitrogen are higher than this, effectively delaying the movement of nutrient to the lake but also making restoration of the lake a very long-term project. Planned action for improving Lake Rotorua water quality includes diverting a spring-fed stream away from the lake and buying up intensively farmed land to change the land use to low input forest (White et al. 2007). These are expensive options that will take many years to implement and for which it will take even longer to see the results.

ESSAY QUESTIONS

1 **Explain the Hjulstrom curve and describe its importance for suspended loading in a river.**

2 **Discuss the importance of the BOD_5 test in the assessment of overall water quality for a river.**

3 **Compare and contrast the direct measurement of water quality parameters to the use of proxy measures for the overall assessment of water quality in a river.**

4 **Explain the major causes of enhanced (cultural) eutrophication in a river system and describe the measures that may be taken to prevent it occurring.**

5 **Explain how residence time of water in a catchment can influence the water quality response to land use change.**

FURTHER READING

Chapman, D. (ed.) (1996) *Water quality assessments: A guide to the use of biota, sediments and water in environmental monitoring* (2nd edition). Chapman & Hall, London.
A comprehensive guide to water-quality assessment.

Dodds, W.K. and Whiles, M.R. (2010) *Freshwater ecology: Concepts and environmental applications of limnology* (2nd edition) Academic Press, San Diego.
A comprehensive introduction to water quality impacts on the aquatic environment.

Gray, N.F. (2010) *Water technology: An introduction for environmental scientists and engineers* (3rd edition) IWA Publishing, London.
An introduction to the engineering approach for controlling water pollution.

11

WATER RESOURCES IN A CHANGING WORLD

LEARNING OBJECTIVES

When you have finished reading this chapter you should have:

- An understanding of water resource management.
- An understanding of the main issues of change that affect hydrology.
- An understanding of how hydrological investigations are carried out to look at issues of change.
- A knowledge of the research literature and main findings in the issues of change.
- A knowledge of case studies looking at change in different regions of the world.

We live in a world that is constantly adjusting to change. This applies from the natural, through to the economic world and is fundamental to the way that we live our lives. The theory of evolution proposes that in order to survive, each species on the planet is changing over a long time period (through natural selection) in order to adapt to its ecosystem fully. Equally, economists would say that people and businesses need to adapt and change to stay competitive in a global economy. If, as was argued in the introductory chapter of this book, water is fundamental to all elements of our life on this planet, then we would expect to see hydrology constantly changing to keep up with our changing world. It is perhaps no great surprise to say that hydrology has, and is, changing – but not in all areas. The principle of uniformitarianism states in its most elegant form: 'the present is the key to the past'. Equally, it could be said that the present is the key to the future and we can recognise this with respect to the fundamentals of hydrology. By the end of the twenty-first century people may be living in a different climate from now, their economic lives may be unlike ours, and almost certainly their knowledge of hydrological processes will be greater than at present. However, the hydrological processes will still be operating in the same manner, although maybe at differing rates to those that we measure today. Koutsoyiannis (2013) points out that hydrology has developed as a science through understanding change at different timescales but that humans can never fully control that change and therefore change and uncertainty are key components for us to understand for the future.

The early chapters of this book have been concerned with hydrological processes and our assessment of them. Our knowledge of the processes will improve, and our methods of measurement and estimation will get better, but the fundamentals will still be the same. In this final chapter several hydrological issues are explored with respect to managing water resources and change that might be expected. The issue of change is explored in a water resource management context: how we respond to changes in patterns of consumption; increasing population pressure and possible changes in climate. The topics discussed here are not exhaustive in covering all issues of change that might be expected in the near future, but they do reflect some of the major concerns. It is meant as an introduction to issues of change and how they affect hydrology; other books and articles cover some of these issues in far more depth (e.g. McDonald and Kay 1988; Acreman 2000; Arnell 2002; Koutsoyiannis 2013). The first broad topic of discussion is water resource management, particularly at the local scale. The second topic is the one that dominates the research literature in natural sciences at present: climate change. The third and fourth topics are concerned with the way we treat our environment and the effect this has on water resources: land use change and groundwater depletion. The final topic is urban hydrology – of great concern, with an ever-increasing urban population all around the world.

WATER RESOURCE MANAGEMENT

When the topic of water resource management is discussed it is often difficult to pinpoint exactly what authors mean by the term. Is it concerned with all aspects of the hydrological cycle or only with those of direct concern to humans, particularly water consumption? As soon as the term 'resource' is introduced then it automatically implies a human dimension. Water is a resource because we need it, and there are ways that we can manipulate its provision, therefore water resource management is a very real proposition. If we are going to manage the water environment is it purely for consumption or are there other uses that need protection and management? Over the past hundred years there has been a large-scale rise in the amount of time people in the western world spend at leisure. Leisure activities include sports such as fishing, canoeing and boating, all of which require clean, fully flowing rivers. Thought is now given to the **amenity value** of rivers and lakes (i.e. how useful they are as places of pleasure without necessarily providing a direct economic return). Management of the water environment needs to be designed to maintain and enhance amenity values. Equally we have an obligation to protect the water environment for future generations and for other species that co-exist with the water. Therefore water resource management needs to embrace sustainable development in its good practices. It is clear that water resource management has to embrace all of these issues and at the same time adapt to changing views on what is required of water management.

Almost all of the processes found in the hydrological cycle can be manipulated in some way. Table 11.1 sets out some of these interventions and the implications of their being dealt with by those involved in water resource management. It is immediately apparent from Table 11.1 that the issues go far beyond the river boundary. For example, land use change has a huge importance for water resource management, so that any decisions on land use need to include consultation with water resource managers. It is important that a legislative framework is in place for countries so that this consultation does take place. Likewise for other areas where human intervention may have a significant impact on water resources. The issue of finding the correct management structures and legislation is investigated in the Case Study looking at how water resources have been managed in England (see pp. 235–238). The changing world in this case has been through increasing population pressures, but probably more importantly is an adaptation to changing political beliefs.

Table 11.1 Manipulation of hydrological processes of concern to water resource management

Hydrological process	*Human intervention*	*Impact*
Precipitation	Cloud seeding	Increase rainfall (?)
Evaporation	Irrigation	Increase evaporation rates
	Change vegetation cover	Alter transpiration and interception rates
	Change rural to urban	Increase evaporation rates
Storage	Change land use	Alter infiltration rates
	Aquifer storage and recovery (ASR)	Manipulating groundwater storage
	Land drainage	Lowering of local water tables
	Building reservoirs	Increasing storage
Runoff	Change land use	Alter overland flow rates
	Land drainage	Rapid runoff
	River transfer schemes	Alter river flow rates
	Water abstraction	Removing river water and groundwater for human consumption

A key part of water resource management involves water allocation: the amount of water made available to users, including both out of stream users (e.g. irrigation, town water supply) and instream environmental use (e.g. amenity values, supporting aquatic species). Water allocation in a resource management context is about how to ensure fair and equitable distribution of the water resource between groups of stakeholders. In South Africa, legislation introduced in 1998 designated that water for minimum human and ecological needs constitutes an untouchable reserve (Jaspers 2001). This promotes human usage and instream ecology above other usages (e.g. agriculture, industry). In the USA the way in which 'water rights' are associated with land property rights means that there are many examples where farms have been bought specifically for the associated water right rather than the agricultural value of the land. This is particularly true in western states of the USA like Colorado, where water is definitely a scarce resource. The city of Boulder has steadily acquired agricultural water rights which it has then used for municipal supply. In the 1990s, Boulder 'gave back' $12M of water rights to ensure continuous flows in Boulder Creek (Howe 1996). This is essentially a reallocation of Boulder Creek water in recognition of aesthetic and environmental needs ahead of human usage.

Case study

CHANGING STRUCTURES OF WATER RESOURCE MANAGEMENT – ENGLAND AS AN EXAMPLE

The major issues of concern for water resource management in England are: water supply; waste disposal; pollution and water quality; and fisheries/aquatic ecosystems management. Other interrelated issues that come into water resource management are flood defence and navigation. Historically it is the first three that have dominated the political agenda in setting up structures to carry out water resource management in England.

History of change

Towards the end of the nineteenth century, great municipal pride was taken in the building of reservoirs to supply water to urban centres in England. At the same time many sewage treatment works were built to treat waste as there was a recognition of the need to keep waste water separate from water supply sources. These were built and run by local councils and replaced a previously haphazard system of private water supply and casual disposal of waste. At this stage, water resource management resided firmly at the local council level. This system continued until the Water Act of 1973 was passed, a bill that caused a major shake-up of water resource management in England and Wales. The major aim of the 1973 act was to introduce holistic water management through administrative boundaries that were governed by river catchments rather than political districts. There was some success in this regard with Regional Water Authorities (RWAs) taking over the water management issues listed above from local councils and other bodies. One of the difficulties with this management structure was the so-called 'poacher–gamekeeper' problem whereby the RWAs were in charge of both waste disposal and pollution control; creating a conflict of interest in water management. Throughout their existence the RWAs operated from a diminishing funding base which led to a lack of investment in waste treatment facilities. It was obvious by the end of the 1980s that a raft of upcoming European Community legislation on water quality would require a huge investment in waste treatment to meet water quality standards. The government at the time decided that this investment was best supplied through the private sector and in 1989 a new Water Act was introduced to privatise the supply of drinking water and wastewater treatment. This has created a set of private water companies with geographic boundaries essentially the same as the RWAs. At the same time a new body, the National Rivers Authority (NRA), was set up to act as a watchdog for water quality. This management structure is still in place in 2017 except that since 1996 the NRA had been subsumed within a larger body, the Environment Agency. The Environment Agency (amongst other duties) monitors river water quality, prosecutes polluters, and issues licenses for water abstraction and treated waste disposal.

In 2000 the European Union (and therefore the United Kingdom) adopted the Water Framework Directive which required the setting of environmental objectives for each of their water bodies (including groundwater) for 2009–2015. This required the Environment Agency to ensure that water bodies complied with the objectives as measured by environmental standards such as pollutant concentrations, health of fish populations, and groundwater quantity.

How has this change affected water resource management?

The answer to this question can be answered by looking at figures for water abstraction and measured water quality over time.

Figure 11.1 shows the water abstracted for supply in England and Wales from 1961 to 2000. During the period of public control (whether councils or RWAs) there was a steady increase in the amount of water abstracted, apart from a blip in the mid-1970s when there were two particularly dry years. Since privatisation in 1989 there has been a flattening and then decline in the amount of water abstracted. This decline cannot be accounted for by the population which has shown a gradual increase during the same time period (see Figure 11.1). There are two causes of this decline: less water being used for industry due to a decline in the industrial sector (although this has been in decline since the early 1980s), and a drop in the amount of leakage from the supply network. This second factor has been forced upon the water companies by political pressure,

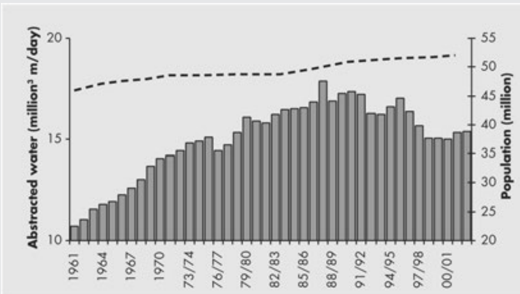


Figure 11.1 Abstracted water for England and Wales 1961–2003 (bar chart) with population for England and Wales 1971–2001 shown as a broken line.

Source: Data from OFWAT and various other sources

particularly following a drought in 1995 and allegations of water supply mismanagement in Yorkshire Water plc. The reduction in leakage has required considerable investment of capital into water supply infrastructure. Overall the water abstracted for public supply is now at the same level as the late 1970s, despite a population rise of nearly 4 million in the corresponding period. The decline has also been achieved despite an increase in the amount of water consumed per household. The United Kingdom has the highest water consumption per capita in Europe and this is rising – a reflection of changing washing habits and an increase in the use of dishwashers. This decline in water abstractions is good for the aquatic environment as it allows a more natural river regime and groundwater system to operate with less human intervention.

It is more difficult to ascertain how the changing management structure has affected water quality in England and Wales because the ways of describing water quality have changed with time. Figure 11.2 shows river water quality assessment using three different scales. The figures shown are achieved by sampling water quality over a period of time (normally years) for hundreds of river reaches around the country. During the first period, when the control was either local council or RWAs, there was very little change when assessed on a four-point scale. After 1980 the scale was changed to five points and during this period of predominantly RWA control, the A grade water quality declined while the percentage of 'fair' water quality river reaches increased. In 1995 the scale was changed again to make it five points and also a differentiation was made between biological and chemical monitoring of water quality (only the chemical results are shown in Figure 11.2). The measurements for 1990 have

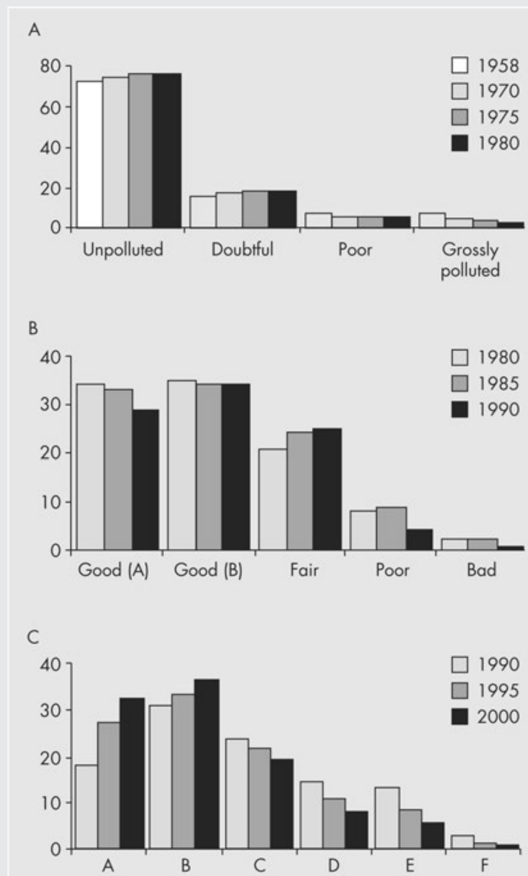


Figure 11.2 Water quality assessment for three periods between 1985 and 2000. An explanation of differing scales is given in the text.

been recalculated onto the 5-point scale to provide some continuity between assessments. Since 1990 there has been a rise in the two highest categories of water quality at the expense of all the others, a response to the extra investment in wastewater treatment provided by the privatised water companies. The percentage figures for 2000 suggest that the lowest category of water quality has been almost eliminated, although when this is recalculated as river length it shows that there are approximately 162 km of extremely poor quality river reaches (out of a total 40,588 km assessed). The 2005 figures show very little change from 2000. In another analysis the Environment Agency figures show that the percentage of rivers in England of 'good' chemical water quality has risen from 55 per cent in 1990 to 80 per cent in 2009. In Wales it rose from 80 per cent to 87 per cent in the same period.

In summary it has to be said that the biggest impact on the water environment for England and Wales has been the privatisation of supply and wastewater treatment, and the setting up of a separate environmental watchdog organisation, in 1989. Prior to this the water quality remained static or worsened and the amount of water abstracted continued to rise. Although the integration of water management into a holistic structure based around the water catchments (i.e. the RWAs) was a noble idea it made very little difference to the crude measures shown here. Since privatisation the water quality has improved and the total amount of water abstracted has decreased. Fundamentally the reason for this is that the investment in infrastructure has risen dramatically since privatisation. It is probably reasonable to surmise that given the same increase in investment, an RWA structure would have seen a similar improvement. It was never likely, in the political climate of the late 1980s, that this investment would have come from the public purse. It has been left to private companies, and more particularly their customers, to pick up the cost of that investment. The example of changing water management structure in England and Wales shows us that this type of change can have significant impacts on the overall hydrology of a region.

Case study

WATER ALLOCATION IN THREE CONTRASTING COUNTRIES

Demands for water vary according to the climate of an area and the traditional uses of water. This is illustrated in Figure 11.3 which contrasts the uses of abstracted water in New Zealand, the United Kingdom and South Korea. These are three countries of similar area but quite different population sizes and water demands. New Zealand has a small population (4.7 million in 2017) with a relatively small industrial sector in an economy dominated by agriculture. The largest water abstractions are from agriculture, the majority of which is used in the spray irrigation of pasture. The United Kingdom has a much higher population (66 million in 2017) with a large industrial sector compared to agriculture. The climate means that irrigation of agricultural land is not common in the UK, hence the small percentage of water used by agriculture compared to industry and household consumption. South Korea has a population of 51 million (in 2017) and large industrial economic sector and yet in Figure 11.3 the water abstraction usages are more similar in profile to New Zealand than the United Kingdom. This is because the predominant agricultural practice used, paddy field rice production, has a very high water usage. So although agriculture is not a large part of the

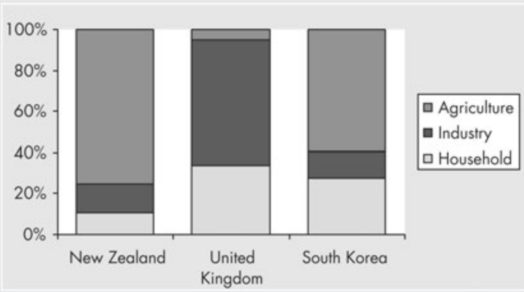


Figure 11.3 Water allocation in three contrasting countries: New Zealand, United Kingdom and South Korea. The figures are broad categories of use for water abstracted in each country.

Source: OECD (2005)

South Korean economy there is a high water demand because of the way it is used.

The predominance of high-tech industry in South Korea also means that industrial water demand is relatively low compared to what might be expected from heavy industry (e.g. steel production). The large populations of South Korea and the UK result in a much higher percentage of total water allocation going to household supply than in New Zealand. It is estimated globally that although rainfall provides about 90 per cent of water used by crops, 70 per cent of all abstracted water is used in irrigation (UNESCO 2006). This suggests that the figures on water used for agriculture in the United Kingdom are unusual and South Korea and New Zealand are closer to the norm.

Water allocation demands do not normally stay static. This is illustrated in Figure 11.4, showing the amount of irrigated land in New Zealand from 1965 to 2017. There has been a doubling of irrigated land approximately every 20 years but even within this period there has been irregular development. Between 1985 and 2002 the majority of irrigation development has occurred since 1995. The increase in irrigation has placed significant stresses on the water system and water allocation regime. Improvements in irrigation technology means that a doubling of the irrigated area does not equate to a doubling in water demand, but it still requires an additional quota of either ground or surface water that is not available for other users, especially instream use.

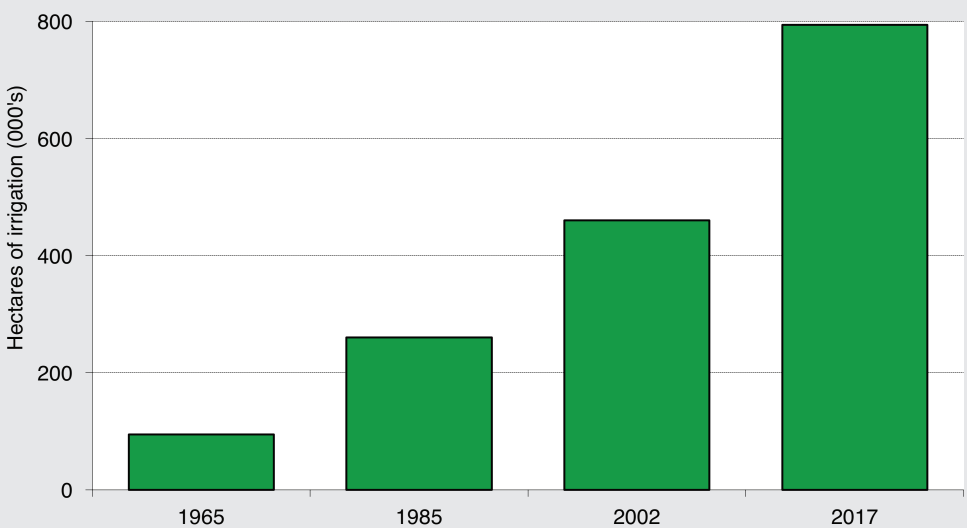


Figure 11.4 Hectares of irrigation in New Zealand from 1965 to 2017.

A key part of water resource management is the involvement of many different sectors of the community in decision-making. This has led to a different approach to water management that stresses integration between different sectors. There are two key concepts in this area: Integrated Water Resource Management (IWRM) and Integrated Catchment Management (ICM). These are both described in more detail below.

Integrated Water Resource Management (IWRM)

The concepts behind IWRM lie in the so-called 'Dublin Principles'. In January 1992, 500 participants, including government-designated experts from 100 countries and representatives of 80 international, intergovernmental and non-governmental organisations attended the International Conference on Water and the Environment in Dublin, Ireland. The conference adopted what has been termed 'the Dublin Statement' which was taken forward to the Earth Summit Conference in Rio de Janiero later that year.

The Dublin Statement established four guiding principles for managing freshwater resources, namely:

1 Fresh water is a finite and vulnerable resource, essential to sustain life, development and the environment.
2 Water development and management should be based on a participatory approach, involving users, planners and policy makers at all levels.
3 Women play a central part in the provision, management and safeguarding of water.
4 Water has an economic value in all its competing uses and should be recognised as an economic good.

These four principles underlie IWRM, especially the concepts of a participatory approach and that water has an economic value (Solanes 1998). An economic good, as used in principle four, is defined in economics as: a physical object or service that has value to people and can be sold for a non-negative price in the marketplace. A major implication from principle four is that water is not a gift or a free right to any water user, it needs to be recognised that using water restricts the usage by others and therefore there is a cost involved in the action.

The Global Water Partnership (www.gwpforum.org) is a leading agency in promoting IWRM. It defines IWRM as:

> a process which promotes the coordinated development and management of water, land and related resources in order to maximise the resultant economic and social welfare in an equitable manner without compromising the sustainability of vital ecosystems.
>
> (Global Water Partnership 2004: 7)

The emphasis within an IWRM approach to water management is on integration between sectors involved in water resources, including local communities (a participatory approach). Although this is promoted as a new approach to resource management it is in many ways a return to traditional values with recognition of the interconnectedness of hydrology, ecology and land management. If there is a large amount of water from a stream allocated to agriculture then there is less available for town water supply and instream ecology. IWRM is a framework for change that recognises this interconnectedness and builds structures to manage water with this in mind. It is an attempt to move away from structures that promote individual sectors competing against each other for the scarce resource of water and moves towards joint ownership of water resource management.

The types of approaches suggested for use within IWRM are illustrated in Table 11.2. These are instruments for change that the Global Water partnership promotes as being integral to IWRM. While these are by no means the only means of achieving an integrated management of water resources they are a useful starting point.

Integrated Catchment Management (ICM)

Integrated Catchment Management (also sometimes referred to as Integrated Water Basin Management, IWBM) is essentially a subset of IWRM. It aims to promote an integrated approach to water and land management but with two subtle differences:

Table 11.2 Eight IWRM instruments for change as promoted by the Global Water Partnership

IWRM instrument for change	*Comments and requirements*
Water resources assessment	Understanding what water resources are available and the water needs of communities. Requires measurements of flows, groundwater levels, etc. and water usage (e.g. metering of take).
IWRM plans	Combining development options, resource use and human interaction. Requires inter-sectoral approach.
Demand management	Using water more efficiently. Requires knowledge of where water losses occur (leakage) and plans on how to promote water efficiency.
Social change instruments	Encouraging a water-oriented society. Requires community education on the importance of using water wisely.
Conflict resolution	Managing disputes, ensuring sharing of water. Requires promotion of trust between sectors and robust dispute settlement systems.
Regulatory instruments	Allocation and water use limits. Requires good knowledge about the amount of available resource and how the hydrological system responds to stress (either natural or human-induced).
Economic instruments	Using value and prices for efficiency and equity. Requires good information on water usage and overall water demand.
Information management and exchange	Improving knowledge for better water management. Requires good data-sharing principles (e.g. between flood control and water supply agencies).

Source: Global Water Partnership (2004)

1 ICM recognises the catchment (or river basin) as the appropriate organising unit for understanding and managing water-related biophysical processes in a context that includes social, economic and political considerations;
2 There is recognition of the spatial context of different management actions and in particular the importance of cumulative effect within a catchment.

By defining a river catchment as the appropriate organising unit for managing biophysical processes there is a recognition that hydrological pathways are important and these provide an appropriate management, as well as biophysical, boundary. Cumulative effect refers to the way in which many small actions may individually have very little impact but when combined the impact may be large. This is true for a river catchment system where individual point discharges of pollution may be small but when accumulated within the river they may be enough to cross an environmental threshold.

Fenemor et al. (2006) have defined the word 'integrated' in an ICM context using three different connotations:

- integration between the local community, science and policy so that the community is linked into the planning and execution of both science and policy and scientific research is being carried out in an environment close-linked into policy requirements and vice versa (see Figure 11.5);
- integration between different scientific and technical disciplines to tackle multi-dimensional problems;
- spatial integration throughout a watershed so that the cumulative impact of different actions can be assessed.

Using this type of definition ICM can be seen as a process that can be used to implement IWRM. One

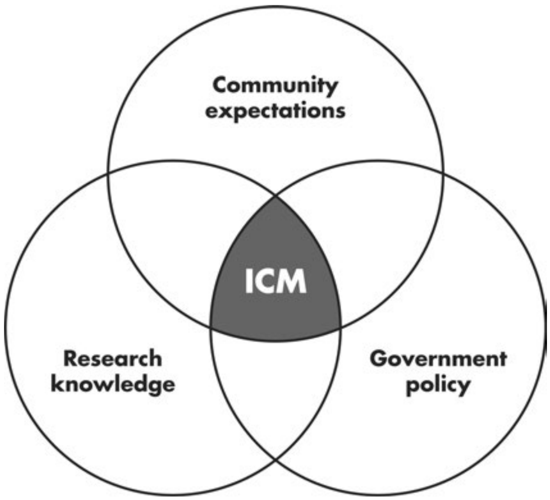


Figure 11.5 The integrating nature of ICM within the context of science, local community and governance.

of the key principles of ICM and IWRM is community involvement through a participatory approach: making sure that everybody can be involved in resource management, not just a few elite within a single organisation. Another key principle of ICM and IWRM is the idea of change. This ranges from extolling change in management structures to cope with modern resource management pressures to making sure the structures can cope with more inevitable changes in the future.

ICM is promoted by UNESCO and the World Meteorological Organisation (WMO) through the Hydrology for the Environment, Life and Policy (HELP) programme. Further details can be found at www.unesco.org/water/ihp/help.

HYDROLOGY AND CHANGE

In water resource management there is a problem concerning the statistical techniques that we use. In frequency analysis there is an inherent assumption that a storm event with similar antecedent conditions, at any time in the streamflow record, will cause the same size of storm. We assume that the hydrological regime is stationary with time. Under conditions of land use or climate change it is quite possible that these conditions will not be met. This makes it difficult to put much faith in a technique such as frequency analysis when it is known that the hydrological regime has changed during the period of record. These are the types of challenges facing water resource management in an ever-changing world. The following section outlines some of the changes possible and uses case studies to demonstrate the possible effects of those changes.

Climate change

At the start of the twenty-first century climate change is the biggest environmental talking point, dominating the scientific media and research agenda. Any unusual weather patterns are linked to the greenhouse effect and its enhancement by humans. The summer of 2006 in Northern Europe was one of the hottest and driest on record and there was drought. At the same time New Zealand experienced one of the wettest winters on record with record snowfalls to sea level, followed by a wet and cold summer. At various times in the media, both these events were linked to global warming. The difficulty with trying to verify any real link to climate change is that hydrological systems naturally contain a huge amount of variability. The extreme events we are experiencing now may be part of that natural variability, or they may be being pushed to further extremes by climate change. However, we do know from analysis of land and sea surface temperatures around the globe that 2016 was the warmest year on record since 1880 and that 16 of the 17 warmest years occurred between 2000 and 2016. The exception to this was 1998 which was the eighth warmest.

Predictions from the Intergovernmental Panel on Climate Change (IPCC 2014) suggest that the earth is likely to experience a global surface temperature rise of 1.5 °C by the end of the twenty-first century. Linked to this prediction is a likely increase in sea level of 26–82 cm and changes in the temporal and spatial patterns of precipitation. The predicted changes in precipitation are not uniform with high latitudes and equatorial Pacific regions seeing an increase in annual precipitation while many mid-latitude and subtropical dry regions will likely decrease (IPCC 2014). The predicted precipitation changes within regions are highly variable depending on changes in the dominant weather patterns for a site. However, IPCC (2014) predict that extreme precipitation events over wet tropical and mid-latitude land masses 'will very likely become more intense and more frequent'. All of these predicted changes will influence the hydrological cycle in some way.

The IPCC predictions on impacts on water resource matters are shown in Table 11.3. At the very simple level a temperature rise would lead to greater evaporation rates, which in turn puts more water into the atmosphere. This may lead to higher precipitation rates, or at least changes in precipitation patterns. How this impacts the hydrology of an individual river catchment is very difficult to predict. The most common method to make predictions is to take the broad-brush predictions from a global circulation model (often at a scale of 1° latitude and longitude per grid square) and downscale it to the local river catchment level. There are several methods used to downscale the data, and Wilby et al. (2000) show that the choice of method used can influence the modelling predictions dramatically. Lauri et al. (2012) show that changes in wet season flows for the Mekong River are predicted to vary between an 11 per cent decrease and a 15 per cent increase depending on which of the global circulation models are used as inputs. The choice of hydrological model used to predict the hydrological changes can also influence results as Jiang et al. (2007) demonstrate for the Dongjiang River in southern China.

At the global scale, Arnell and Gosling (2013) simulated future climate change predictions across different hydrological domains. They show that under one climate model scenario 47 per cent of the land surface (excluding the polar regions) showed an increase in runoff and 36 per cent a significant decrease. Adam et al. (2009) use knowledge of recent changes in snowmelt hydrology in the western USA to predict global change in snowmelt characteristics. Adam et al. (2009) predict 'strong decreases in winter snow accumulation and spring snowmelt' in mid-latitudes. They particularly point to changes in snowmelt-derived runoff being largest in areas that currently experience large snowfalls.

Global scale studies give broad statements of change, but they require downscaling from global circulation models to a regional or catchment scale – this provides the best information for water management. There have been many studies doing this. In Europe Middelkoop et al. (2001) predict higher winter discharges on the Rhine from 'intensified snow melt and increased winter precipitation'. Arnell and Reynard (1996) used models of river flow to try and predict the effects of differing climate change predictions on the river flows in 21 river catchments in Great Britain. Their results suggest a change in the seasonality of flow and also considerable regional variation. Both these changes are by and large driven by differences in precipitation. The north-west of England is predicted to become wetter while the south-east becomes drier. Overall it is predicted that winters will be wetter and summers drier. This may place a great strain on the water resources for south-east England where by far the greater percentage of people live. In a more recent study Arnell and Reynard (2000) have suggested that flow duration curves are likely to become steeper, reflecting a greater variability in flow. They also predict an increase in flood magnitudes that in the case of the Thames and River Severn have 'a much greater effect than realistic land use change' (Arnell and Reynard 2000: 21). These changes in river flow regime have important implications for water resource management in the future.

In North America Wilby and Dettinger (2000) predict higher winter flows for three river basins in the Sierra Nevada (California, USA). These higher winter flows reflect changes in the winter snowpack due to a predicted rise in both precipitation and temperature for the region. Ficklin et al. (2012) used the Soil and Water Assessment Tool (SWAT) model to assess hydrological

Table 11.3 Predicted impacts of climate change on water resource management area

Water resource area	*Predicted impact*
River flows	There is a high level of confidence that by the mid-twenty-first century, annual average river runoff and water availability are projected to increase by 10–40 % at high latitudes and in some wet tropical areas, and decrease by 10–30 % over some dry regions at mid-latitudes and in the dry tropics, some of which are presently water stressed areas.
Hydrological extremes (floods and droughts)	There is a high level of confidence that drought-affected areas will increase in extent. Heavy precipitation events, which are very likely to increase in frequency, will augment flood risk.
Snow and ice cover	There is a high level of confidence that over the course of the twenty-first century, water supplies stored in glaciers and snow cover will decline.
Forest production	There is medium confidence that globally, commercial timber productivity will rise modestly with climate change in the short to medium term, with large regional variability around the global trend. This may affect water yield from forestry-covered catchments.
Coastal flooding	There is very high confidence that many millions more people will be flooded every year due to sea-level rise by the 2080s. The numbers affected will be largest in the mega-deltas of Asia and Africa, while small islands are especially vulnerable.

Source: IPCC (2007)

impacts from downscaled climate change predictions for the Mono Lake Basin in eastern California. They predict a 15 per cent decrease in average annual streamflow; less wet hydrological years and more frequent droughts. Further north, Morrison et al. (2002) have investigated riverflow and temperature projections in the Fraser River, British Columbia, Canada. Their modelling work predicts small average flow increases but peak flows, which are snowmelt derived, are predicted to decrease by about 18 per cent and occur earlier in the year. Morrison et al. (2002) also predict an increase in summer water temperatures which has implications for salmon spawning in the catchment.

In non-temperate regions of the world predictions vary on climate change. Parry (1990) suggests rainfall in the Sahel region of Africa will stay at current levels or possibly decline by 5–10 per cent. Parry (1990) also suggests a 5–10 per cent increase in rainfall for Australia, although this may have little effect on streamflow when linked with increased evaporation from a 2 °C temperature rise. Chiew et al. (1995) highlight the large regional variations in predictions of hydrological change in Australia. The wet tropical regions of north-east Australia are predicted to have an increase in annual runoff by up to 25 per cent, Tasmania a 10 per cent increase, and a 35 per cent decrease in South Australia, by 2030. The uncertainty of this type of prediction is illustrated by south-east Australia where there are possible runoff changes of ±20 per cent (Chiew et al. 1995). Similarly, Kaleris et al. (2001) attempted to model the impacts of future climate change on rivers in Greece but concluded that the 'error of the model is significantly larger than climate change impacts' and therefore no firm conclusions could be made. Lauri et al. (2012) have modelled the Mekong River in South-east Asia under climate change scenarios and concluded that the impacts of climate change are likely to be less than that seen in reservoir management for hydropower. This is a result of a wide range of predicted changes in hydrology (e.g. – 11 per cent to +15 per cent change in wet season flows) but also that the scale of human interventions on the Mekong are substantial.

Overall the changes in hydrology under climate change scenarios vary widely. This doesn't mean that climate change should be ignored, quite the opposite in fact, it stresses the importance of understanding the hydrology at a local catchment scale and further downscaling studies at that scale. Of particular concern in water resource management are the predictions for increased rainfall intensity in many places which will undoubtedly lead to increased flooding risk.

Change in land use

The implications of land use change for hydrology has been an area of intense interest to research hydrologists over the last 50 or more years. Issues of land use change affecting hydrology include increasing urbanisation (see pp. 251–253), changing vegetation cover, land drainage and changing agricultural practices leading to salination.

Vegetation change

In Chapter 4, a Case Study showed the effect that trees have on evaporation and interception rates. This is a hydrological impact of vegetation cover change, a subject that Bosch and Hewlett (1982) review in considerable depth. In general, Bosch and Hewlett conclude that the greater the amount of deforestation the larger the subsequent streamflows will be, but the actual amount is dependent on the vegetation type and precipitation amount. This is illustrated by the data in Table 11.4. In the Australian study of Crockford and Richardson (1990) the large range of values are from different size storms. The high interception losses were experienced during small rainfall events and vice versa. The interception loss from the Amazonian rain forest is remarkably low, reflecting a high rainfall intensity and high humidity levels. Overall there is a high degree of variability in the amount of interception loss that is likely to occur. While it may be possible to say that in general a land use change that has increased tree cover will lead to a water loss, it is not easy to predict by how much that will be.

Table 11.4 The amount of interception loss (or similar – see note below) for various canopies as detected in several studies

Canopy cover	*Interception loss*	*Source*
Eucalypt forest (Australia)	5–26 % per rainfall event	Crockford and Richardson (1990)
Pine forest (Australia)	6–52 % per event	Crockford and Richardson (1990)
Oak stand (Denmark)	15 % of summer rainfall	Rasmussen and Rasmussen (1984)
Amazonian rainforest	9 % of annual	Lloyd et al. (1988)
Sitka spruce (Lancashire, UK)	38 % of annual precipitation*	Law (1956)
Sitka spruce (Wales)	30 % of annual precipitation*	Kirby et al. (1991)
Grassland (Wales)	18 % of annual precipitation*	Kirby et al. (1991)
Young Douglas fir (New Zealand) (closed canopy)	27 % of 7-month summer/ autumn period	Fahey et al. (2001)
Mature Douglas fir (NZ)	24 % of 7-month summer/ autumn period	Fahey et al. (2001)
Young *Pinus radiata* (NZ) (closed canopy)	19 % of 7-month summer/ autumn period	Fahey et al. (2001)

Note: *The figures denoted with an asterisk are actually evapotranspiration values rather than absolute interception loss, leading to higher values.

Fahey and Jackson (1997) conclude that with the loss of forest cover, both low flows and peak flows increase. The low flow response is altered primarily through the increase in water infiltrating to groundwater without interception by a forest canopy. The peak flow response is a result of a generally wetter soil and a low interception loss during a storm when there is no forest canopy cover. The time to peak flow may also be affected, with a more sluggish response in a catchment with trees. In a modelling study, Davie (1996) has suggested that any changes in peak flow that result from afforestation are not gradual but highly dependent on the timing of canopy closure.

In Chapter 8 the issue of measurement scale was discussed, and it is particularly pertinent for issues of land use change. There has been considerable debate in the hydrological research literature as to how detectable the effects of deforestation are in large river catchments. Jones and Grant (1996) and Jones (2000) analysed data from a series of paired catchment studies in Oregon, USA and concluded that there was clear evidence of changes in interception rates and peak discharges. Thomas and Megahan (1998) reanalysed the data used by Jones and Grant (1996) and came to the conclusion that although there was clear evidence of changes in peak flows in the small-scale catchment pairs (60–100 km^2) there was no change, or inconclusive evidence for change, in the large catchments (up to 600 km^2). There has followed a series of letters between the authors disputing various aspects of the studies (see *Water Resources Research* **37**: 175–183). This debate in the research literature mirrors the overall concern in hydrology over the scale issue. There are many processes that we measure at the small hillslope level that may not be important when scaled up to larger catchments.

Land drainage

Land drainage is a common agricultural 'improvement' technique in areas of high rainfall and poor natural drainage. In an area such as the Fens of Cambridgeshire, Norfolk and Lincolnshire in England this has taken the form of drains or canals and an elaborate pumping system, so that the natural wetlands have been drained completely. The result of this has been the utilisation of the area for intensive agricultural production since the drainage took place in the seventeenth and eighteenth centuries. Since that time the land has sunk, due to the removal of water from the peat-based soils, and the area is totally dependent on the pumping network for flood protection. To maintain this network vegetation control and clearance of silt within channels is required, a cost that can be challenged in terms of the overall benefit to the community (Dunderdale and Morris 1996).

At a smaller scale, land drainage may be undertaken by farmers to improve the drainage of soils. This is a common practice throughout temperate regions and allows soils to remain relatively dry during the winter and early spring. The most common method of achieving this is through a series of tile drains laid across a field that drain directly into a water course (often a ditch). Traditionally tile drains were clay pipes that allowed water to drain into them through the strong hydraulic gradient created by their easy drainage towards the ditch. Modern tile drains are plastic pipes with many small holes to allow water into them. Tile drains are normally laid at about 60 cm depth and should last for at least 50 years or more. To complement the tile drains, **mole drainage** is carried out. This involves dragging a large, torpedo-shaped metal 'mole' behind a tractor in lines orthogonal to the tile drains. This creates hydrological pathways, at 40–50 cm depth, towards the tile drains. Mole draining may be a regular agricultural activity, sometimes every 2 to 5 years in heavy agricultural land (i.e. clay soils). Normal plough depth is around 30 cm, so that the effects of mole draining last beyond a single season.

The aim of tile and mole drainage is to hold less water in a soil. This may have two effects on the overall hydrology. It allows rapid drainage from the field, therefore increasing the flashy response (i.e. rapid rise and fall of hydrograph limbs) in a river. At the same time the lack of soil moisture may lead to greater infiltration levels and hence less overland flow. Spaling (1995) noted that in southern Ontario, Canada, land drainage alters timing and volume of water flow at the field scale, but it is difficult to detect this at the watershed scale. Hiscock et al. (2001) analysed 60 years of flow records for three catchments in Norfolk, UK to try and detect any change in the rainfall–runoff relationship during this time. The conclusion of their study was that despite much land drainage during the period of study, the rainfall–runoff relationship 'remained essentially unchanged' (Hiscock et al. 2001). This lack of change in overall hydrology, despite the known land drainage, may be due to the two hydrological impacts cancelling each other out, or else that the impact of land drainage is small, particularly at the large catchment scale.

Land drainage can be a significant factor in upland areas used for forestry. A common technique in Europe is using the plough and furrow method of drainage. A large plough creates drains in an area, with the seedlings being planted on top of the soil displaced by the plough (i.e. immediately adjacent to the drain but raised above the water table). Like all land drainage this will lower the water table and allow rapid routing of stormflow. A study at a small upland catchment in the north-east of England has shown that land drainage effects are drastic, and only after 30 years of afforestation has the impact lessened (Robinson 1998). This long recovery time may be a reflection of the harsh environment the trees are growing in; other areas have recovered much faster.

Salination

Salination is an agricultural production problem that results from a build-up of salt compounds in the surface soil. Water flowing down a river is almost never 'pure', it will contain dissolved solids in the form of salt compounds. These salt compounds are derived from natural sources such as the weathering of surface minerals and sea spray contained in rainfall. When water evaporates the salts are left behind, something we are familiar with from salt lakes such as in Utah, central Australia, and the Dead Sea in the Middle East. The same process leads to salinity in the oceans.

Salination of soils (often also referred to as salinisation) occurs when there is an excess of salt-rich water that can be evaporated from a soil. The classic situation for this is where river-fed irrigation water is used to boost agricultural production in a hot, dry climate. The evapotranspiration of salt-rich irrigation water leads to salt compounds accumulating in the soil, which in turn may lead to a loss of agricultural production as many plants fail to thrive in a salt-rich environment. Although salination is fundamentally an agronomic problem it is driven by hydrological factors (e.g. water quality and evaporation rates), hence the inclusion in a hydrological textbook.

Salination of soils and water resources have been reported from many places around the world. O'Hara (1997) provides data on waterlogging and subsequent salination in Turkmenistan, the direct result of irrigation. Gupta and Abrol (2000) describe the salinity changes that have occurred in the Indo-Gangetic Plains on the Indian subcontinent following increased rice and wheat production. Flugel (1993) provides data on irrigation return flow leading to salination of a river in the Western Cape Province of South Africa. Prichard et al. (1983) report salination of soils from irrigation in California, USA. Irrigation water often has a high total dissolved solids (TDS) load before being used for agricultural production. Postel (1993) suggests that typical values range from 200–500 mg/l, where water is considered brackish at levels greater than 300 mg/l. Postel also states that using this type of irrigation water under a normal irrigation level would add between 2 and 5 tons of salt per hectare per year. The vast majority of this salt is washed out of the soil and continues into a water table or river system; some, though, will be retained to increase salination in the soil.

Case study

SALINATION OF WATERWAYS IN AUSTRALIA: A SALINITY PROBLEM FROM LAND USE CHANGE

Salination of surface waters is a huge problem for large areas of Australia. Sadler and Williams (1981) estimate that a third of surface water resources in south-west Western Australia can be defined as brackish (from Williamson et al. 1987). In the same region it is estimated that 1.8 million hectares of agricultural land is affected by salinity problems (Nulsen and McConnell 2000). In an assessment of ten catchments in New South Wales and Victoria (total land area of 35.7 million hectares) it is estimated that 4.1 per cent of the land area is affected by salinity and that this imposes a cost of $122 million (Australian dollars) on agricultural production (Ivey ATP 2000).

This salinity problem is a steady increase in concentration of salt compounds in rivers, leading to the surface water becoming unusable for public supply or irrigation. In south-west Western Australia salinity levels in soils are typically between 20 and 120 kg/m^2 (Schofield 1989 and Williamson et al. 1987 quote a range of 0.2–200 kg/m^2). This is very high and is a result of thousands of years of low rainfall and high evaporation leading to an accumulation of wind-borne sea salt in the soil. The natural vegetation for this area is deep-rooted eucalypt forest and savannah woodland which has a degree of salt tolerance and ability to extract water from deep within the soil. The ability to draw water from deep within a soil maintains a high soil water deficit which is filled by seasonal rainfall (Walker et al. 1990). The removal of this native vegetation and replacement with shallow rooted crops (particularly wheat) and pasture has led to a fundamental change in the hydrology, which in turn has led to a change in salinity. The replacement vegetation does not use as much water, leading to greater levels of groundwater recharge and rising water tables. This is particularly so for wheat which is largely dormant, or when the ground is fallow, during the wetter winter season. The groundwater is often saline and in addition to this, as the water table rises it takes up the salt stored in soils. Rivers receive more groundwater recharge (with saline water) and the streams increase in salinity. The link between vegetation change and increasing salinity levels was first proposed by Wood (1924) and has since been demonstrated through field studies.

Williamson et al. (1987) carried out a catchment-based study of vegetation change and increasing salinity in Western Australia. The study monitored salinity and water quantity in four small catchments, two of which were cleared of native vegetation and two kept as controls. The monitoring took place between 1974 and 1983 with the vegetation change occurring at the end of 1976 and start of 1977; a selection of results is shown here. There was a marked change in the hydrological regime (see Figure 11.6), with a large increase in the amount of streamflow as a percentage of rainfall received. This reinforces the idea of Wood (1924) that the native vegetation uses more water than the introduced pasture species.

In terms of salinity there was also a marked change although this is not immediately evident from a time series plot (Figure 11.7). The chloride

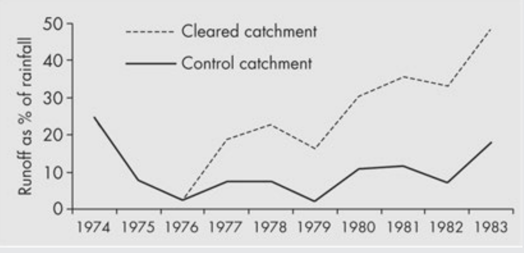


Figure 11.6 Streamflow expressed as a percentage of rainfall for two catchments in south-west Western Australia. The control maintained a natural vegetation while in the other catchment the bush was cleared during 1976/77 and replaced with pasture.

Source: Data from Williamson et al. (1987)

concentration in streamflow is a good indicator of salinity as it is one of the main salts that would be expected to be deposited from sea spray, however it is not the total salinity. In Figure 11.7 there appears to be an increasing difference between chloride concentrations with time. Chloride concentration shows considerable variation between years which is related to variation in rainfall between years. The peaks in salinity correspond to years with high rainfall. To remove this factor Williamson et al. (1987) calculated the chloride concentration as a ratio between output (measured in the streamflow) and input (measured in the rainfall). This is shown in Figure 11.8.

When the chloride level is expressed as this output/input ratio (Figure 11.8) it is easy to see a marked difference following the vegetation change. In the years following 1976/77 there is considerably more output of chloride than input (i.e. the ratio is well above a value of 1), a result of the chloride being leached out of the soil. In this manner the chloride concentration in the river is staying at a high level even when there is a low input (i.e. low rainfall). Prior to vegetation change the ratio is approximately even, the chloride inputs and output had reached some type of equilibrium. Given enough time the same would happen again with the new vegetation cover, but first a large store of chloride would be released from the soil. This is a case where the vegetation change has upset the hydrological balance of a catchment, which in turn has implications for water quality.

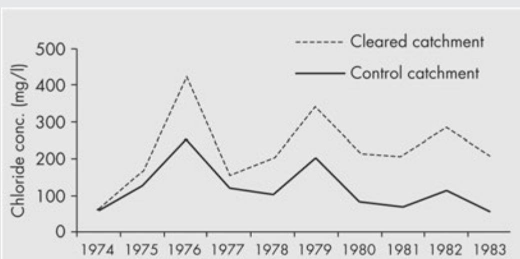


Figure 11.7 Chloride concentrations for two catchments in south-west Western Australia. These are the same two catchments as in Figure 11.6. NB World Health Organisation guidelines suggest that drinking water should have a chloride concentration of less than 250 mg/l.

Source: Data from Williamson et al. (1987)

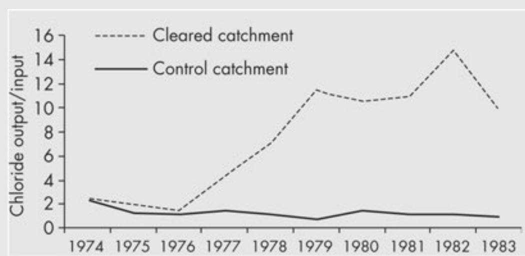


Figure 11.8 Chloride output/input ratio for two catchments in south-west Western Australia. These are the same two catchments as in Figures 11.6 and 11.7. Input has been measured through chloride concentrations in rainfall while output is streamflow.

Source: Data from Williamson et al. (1987)

Groundwater depletion

In many parts of the world there is heavy reliance on aquifers for provision of water to a population. In England around 30 per cent of reticulated water comes from groundwater, but that rises to closer to 75 per cent in parts of south-east England. The water is extracted from a chalk aquifer that by and large receives a significant recharge during the winter months. Apart from very dry periods (e.g. the early 1990s) there is normally enough recharge to sustain withdrawals.

Not all groundwater is recharged so readily. Many aquifers have built up their water reserves over millions of years and receive very little infiltrating rainfall on a year by year basis. Much of the Saudi peninsula in the Middle East is underlain by such an aquifer. The use of this water at high rates may lead to groundwater depletion, a serious long-term problem for water management. The Ogallala aquifer Case Study introduces groundwater depletion problems in the High Plains region of the USA.

Case study

OGALLALA AQUIFER DEPLETION

The Ogallala aquifer (also called the High Plains aquifer) is a huge groundwater reserve underlying an area of approximately 583,000 km^2 in the Great Plains region of the USA. It stretches from South Dakota to Texas and also underlies parts of Nebraska, Wyoming, Colorado, Kansas and New Mexico (see Figure 11.9).

The aquifer formed through erosion from the Rocky Mountains to its west. The porous material deposited from this erosion was filled with water from rivers draining the mountains and crossing the alluvial plains. This has created a water reserve that in places is 300 m deep. A major problem is that now the aquifer is isolated from the Rocky Mountains as a recharge source and has to rely on natural replenishment from local rainfall and infiltration. This is a region that receives around 380–500 mm of rainfall per annum and has very high evaporation rates during the summer. The climate is classified as semi-arid.

Ever since Europeans first settled on the Great Plains, the Ogallala aquifer has been an important water source for irrigation and drinking water supply. Since the 1940s there has been rapid expansion in the amount of irrigated land in the region (see Figure 11.10) so that in 1990 as much as 95 per cent of water drawn from the aquifer was used for irrigating agricultural land (McGuire and Fischer 1999). Improving technology has meant that the windmill driven irrigation that was predominant in the 1940s and 1950s has been replaced with pumps capable of extracting vast amounts of water at a rapid rate. The result of this has been drastic declines in water tables, as much as 30 m in parts of Texas, New Mexico and Kansas (McGuire and Fischer 1999).

There have been various efforts made to reduce the depletion of the Ogallala aquifer but it is made difficult by the importance this area has for agricultural production in the USA. Systems of irrigation scheduling have been introduced to

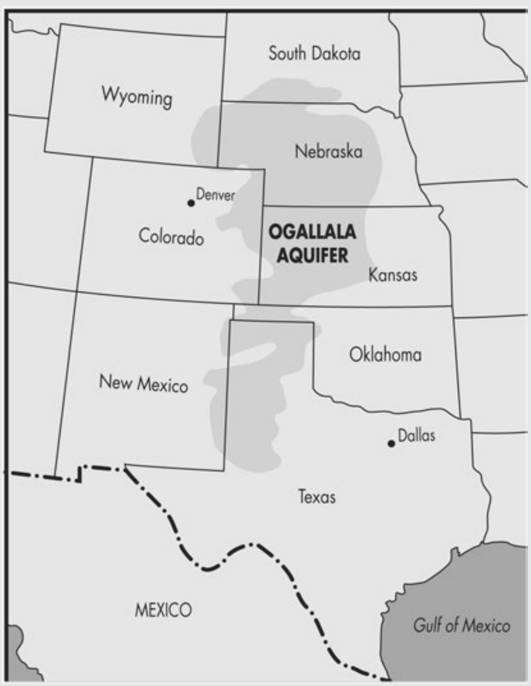


Figure 11.9 Location of the Ogallala aquifer in the Midwest of the USA.

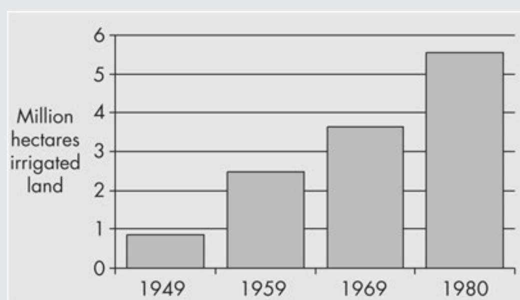


Figure 11.10 Amount of irrigated land using groundwater in the High Plains region.

Source: Data from McGuire (2017)

make the use of irrigated water more efficient. This involves a close monitoring of soil moisture content so that water is only applied when needed by plants and the actual amount required can be calculated. Another management tool to lessen depletion is changing agricultural production so that water thirsty plants such as cotton are not grown in areas that rely on groundwater for irrigation.

The United States Geological Survey (USGS) have been monitoring changes in winter water level in over 8,000 wells since the late 1980s in order to assess the rate of overall groundwater depletion. Average figures for States from before development (about 1950) until 2015 are shown in Figure 11.11. This shows an overall decline in water tables of around 4.8 m but Texas has an average decline of 12.5 m and South Dakota a very small rise (0.15 m). In preparing the second edition of this book there had been a slowing in the rate of water level decline. This was partly attributed to a wet period from 1980 to 1997; however since then there has been substantial drought in the region and water levels have continued to decline. It is estimated that in total around 273 km^3 of groundwater has been depleted from the Ogallala aquifer between 1950 and 2015; this represents approximately 9 per cent of the available stored volume.

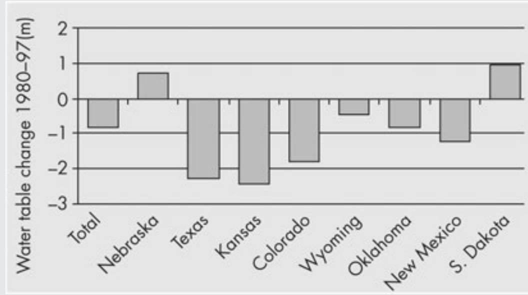


Figure 11.11 Average changes in the water table for states underlying the Ogallala aquifer.

A 9 per cent drop in storage does not sound too extreme but in parts of northern Texas and southern Kansas it is a far higher percentage and represents an unsustainable depletion of the groundwater. It is difficult to see how the decline could be halted without a complete change in agricultural production for the region, but this is unlikely to occur until the price of extracting the water is too high to be economically viable. At the moment the region is using an unsustainable management practice that has led to substantial groundwater depletion and is likely to continue into the near future.

Urbanisation

Many aspects of urban hydrology have already been covered, especially with respect to water quality (see Chapter 10), but the continuing rise in urban population around the world makes it an important issue to consider under the title of change. There is no question that urban expansion has a significant effect on the hydrology of any river draining the area. Initially this may be due to climate alterations affecting parts of the hydrological cycle. The most obvious hydrological impact is on the runoff hydrology, but other areas where urbanisation may have an impact are point source and diffuse pollution affecting water quality, river channelisation to control flooding, increased snow melt from urban areas and river flow changes from sewage treatment.

Urban climate change

In Table 11.5 some of the climatic changes due to urbanisation are expressed as a ratio between the urban and rural environments. This suggests that within a city there is a 15 per cent reduction in the amount of solar radiation reaching a horizontal surface, a factor that will influence the evaporation rate. Studies have also found that the precipitation levels in an urban environment are higher by as much as 10 per cent. Atkinson (1979) detected an increase in summer thunderstorms over London which was attributed to extra convection and

Table 11.5 Difference in climatic variables between urban and rural environments

Climatic variable	*Ratio of city: environs*
Solar radiation on horizontal surfaces	0.85
UV radiation: summer	0.95
UV radiation: winter	0.70
Annual mean relative humidity	0.94
Annual mean wind speed	0.75
Speed of extreme wind gusts	0.85
Frequency of calms	1.15
Frequency and amount of cloudiness	1.10
Frequency of fog: summer	1.30
Frequency of fog: winter	2.00
Annual precipitation	1.10
Days with less than 5mm precipitation	1.10

Source: From Lowry (1967)

condensation nuclei being available. Other factors greatly affected by urbanisation are winter fog (doubled) and winter ultraviolet radiation (reduced by 30 per cent).

Urban runoff change

The changes in climate are relatively minor compared to the impact that impermeable surfaces in the urban environment have on runoff hydrology. Roofs, pavements, roads, parking lots and other impermeable surfaces have extremely low infiltration characteristics, consequently Hortonian overland flow readily occurs. These surfaces are frequently linked to gutters and stormwater drains to remove the runoff rapidly. The result of this is far greater runoff and the time to peak discharge being reduced. Cherkauer (1975) compared two small catchments in Wisconsin, USA. The rural catchment had 94 per cent undeveloped land while the urban catchment had 65 per cent urban coverage. During a large storm in October 1974 (22 mm of rain in 5 hours) the peak discharge from the urban catchment area was over 250 times that of the rural catchment (Cherkauer 1975). The storm hydrograph from this event was considerably more flashy for the urban catchment (i.e. it had a shorter, sharper peak on the hydrograph).

Rose and Peters (2001) analysed a long period of streamflow data (1958–96) to detect differences between urbanised and rural catchments near Atlanta, Georgia, USA. The stormflow peaks for large storms were between 30 and 100 per cent larger in the urbanised catchment, with a considerably shorter recession limb of the hydrograph. In contrast to the stormflows, low flows were 25–35 per cent less in the urban catchment, suggesting a lower rainfall infiltration rate. Overall there was no detectable difference in the annual runoff coefficient (runoff as percentage of precipitation) between urban and rural catchments.

Figure 11.12 shows some data from a steadily urbanising catchment (13 km^2) in Auckland, New Zealand. There has been a drop in the percentage of baseflow leaving the catchment (the baseflow index – BFI) which could be attributed to declining infiltration to groundwater and therefore less water released during the low flow periods. Care needs to be taken in interpreting a diagram like Figure 11.12 because it is also possible that the decline in BFI was caused by an increase in the

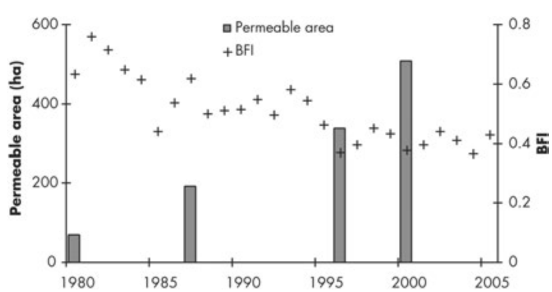


Figure 11.12 Baseflow index (BFI – proportion of annual streamflow as baseflow) with time in a small catchment in Auckland, New Zealand where there has been steady urbanisation. The vertical bars show area of permeable surfaces estimated from aerial photographs at 4 times.

Source: Data courtesy of Auckland Regional Council stormflow and total flow and the actual amount of baseflow has stayed the same. Whichever way, there has been a change in the hydrological regime that can probably be attributed to the rise in permeable surfaces as a result of urbanisation.

Pollution from urban runoff

There is a huge amount of research and literature on the impacts of urbanisation on urban water quality. Davis et al. (2001) link the accumulation of heavy metals in river sediments to urban runoff, particularly from roads. Specific sources are tyre wear and vehicle brakes for zinc, and buildings for lead, copper, cadmium and zinc (Davis et al. 2001). In Paris, Gromaire-Mertz et al. (1999) found high concentrations of heavy metals in runoff from roofs, while street runoff had a high suspended solids and hydrocarbon load. The hydrocarbons are of particular concern, especially the carcinogenic polycyclic aromatic hydrocarbons (PAH) derived from petrol engines. Krein and Schorer (2000) trace PAHs from road runoff into river sediments where they bind onto fine sand and silt particles.

The nature of urban runoff (low infiltration and rapid movement of water) concentrates the pollutants in the first flush of water. Studies have shown that over 80 per cent of pollutant particles are washed into a drainage system within the first 6–10 mm of rain falling (D'Arcy et al. 1998), and often from a very small collection area within the urban catchment (Lee and Bang 2000). This information is important when proposing strategies to deal with the urban pollutant runoff. One of the main methods is to create an artificial wetland within an urban setting so that the initial flush of storm runoff is collected, slowed down, and pollutants can be modified by biological action. Shutes (2001) has a review of artificial wetlands in Hong Kong, Malaysia and England and discussed the role of plants in improving water quality. Scholes et al. (1999) showed that two artificial wetlands in London, England were efficient in removing heavy metals and lowering the BOD of urban runoff during storm events. Carapeto and Purchase (2000) reported similar efficiency for the removal of cadmium and lead from urban runoff.

River channelisation

It is a common practice to channelise rivers as they pass through urban areas in an attempt to lessen floods in the urban environment. Frequently, although not always, this will involve straightening a river reach and this has impacts on the streamflow. Simons and Senturk (1977) list some of the hydrological impacts of channel straightening: higher velocities in the channel; increased sediment transport and possible base degradation; increased stormflow stage (height); and deposition of material downstream of the straightening. The impact of urban channelisation is not restricted to the channelised zone itself. The rapid movement of water through a channelised reach will increase the velocity, and may increase the magnitude, of a flood wave travelling downstream. Deposition of sediment downstream from the channelised section may leave the area prone to flooding through a raised river bed.

Case study

THE CHEONGGYECHEON RIVER: A POST-MODERN RIVER RESTORATION PROJECT?

The Cheonggyecheon river (sometimes spelled as Cheong Gye Cheong) is a tributary of the Han river, the main river flowing through Seoul, South Korea. It has been the centre of historical Seoul since the fourteenth century, originally marking a political boundary and being a source of drinking and washing water. The original name for the river was Gaecheoni, which means 'open stream' in Korean; the current name Cheonggyecheon literally means 'stream'. Flooding of the Cheonggyecheon has always been a problem for Seoul, requiring regular dredging and providing a major hazard for the city. In the eighteenth century 200,000 people were used to dredge the stream, and to widen and strengthen its banks. In the first half of the twentieth century rural depopulation and the migration of people to Seoul led to the river banks becoming overcrowded, with the river acting as a sewer, and there were serious problems with flooding and disease. The Japanese colonial government drew up plans to cover the river and work began in 1937. The covering wasn't completed until 1961, by which time South Korea was an independent state. Between 1967 and 1971 a major expressway was built over the river and the river was essentially forgotten (Figure 11.13). Following concerns in the late 1990s over the safety of the expressway structure it was decided to remove the road and concrete covering and establish the Cheonggyecheon as a riverside recreation area. This was part of an overall Seoul urban revitalisation project promoted by the Mayor Lee Myung-bak.

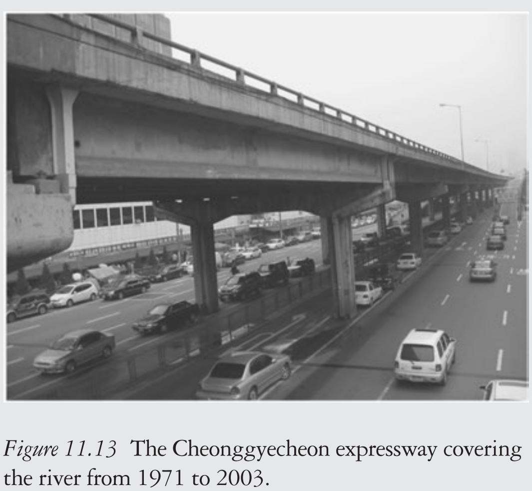


Figure 11.13 The Cheonggyecheon expressway covering the river from 1971 to 2003.

Source: Photo courtesy of Seoul Metropolitan City archive

The term 'restoration' is difficult to apply to the Cheonggyecheon project as the end result is a river far from its original, natural state. However the new Cheonggyecheon fulfils a vital function in modern Seoul: providing open space in a landscape dominated by skyscraper buildings, providing a wildlife corridor, and being a cross between an exhibition area for street sculptures and a civic park (Figure 11.14). The result is a spectacular transformation that has led to much revitalisation of the surrounding district.

The renovation of the river is estimated to have cost US$360 million, three times an early cost

Figure 11.14 The Cheonggyecheon river in a 'restored' state, 2006.

Source: Photo courtesy of Seoul Metropolitan City archive estimation (Lee 2003). In addition to removing the expressway and concrete cover, the design incorporated new roads, a sewerage and stormwater pipe network and had to allow for flooding. The schematic design is shown in Figure 11.15 which shows how the extra flood capacity has been incorporated beneath the roads, the sewerage and stormwater pipe infrastructure being buried within the stream banks. The water for the river is pumped from the nearby Han river via a water treatment plant. This supplies enough water to maintain an average stream depth of 40 cm. The river contains many artificial weirs and fountains, particularly in its upper reaches, ensuring the water oxygen content stays high. Further downstream there is good enough water quality for invertebrates and fish to establish permanent populations. A survey on the ecology of the stream conducted between March and April 2007 by the Seoul Metropolitan Facilities Management Corporation found 30 species of bird and 13 species of fish inhabiting the stream environment. In addition to the aesthetic benefit of having a river within a city centre the ambient air temperature adjacent to the stream has dropped by around 2 °C and the removal of the expressway has led to a decrease in particulate air pollution. However, the water flowing down the river is around 1.4 m^3/s of predominantly treated drinking water, making it an expensive stream to maintain.

Figure 11.15 Schematic diagram of Cheonggyecheon restoration project, showing infrastructure as well as the river.

Source: Image courtesy of Seoul Metropolitan City archive and Lee (2003)

A visit to the Cheonggyecheon river in the evening provides an interesting insight into the way an urban stream can be used as a civic park. The upper reaches of the river are ablaze with multi-coloured neon lights lighting up the stream, the sculptures along the bank and stepping stones across the stream make it a popular place for people to congregate and walk. The Cheonggyecheon has been transformed from a modernist vision (channelised and covered over) into a post-modern concept where the aesthetic function as a civic park is as important as the biophysical function of the river itself.

More details on the Cheonggyecheon Restoration Project can be found at http://english.seoul.go.kr/cheonggye.

Urban snow melt

The influence of urbanisation on snow melt is complicated. Semandi-Davies (1998) suggests that melt intensities are generally increased in an urban area, although shading may reduce melt in some areas. Overall there is a greater volume of water in the early thaw from an urban area when compared to a rural area (Taylor and Roth 1979; Semandi-Davies 1998). This may be complicated by snow clearance operations, particularly if the cleared snow is placed in storage areas for later melting (Jones 1997). In this case the greater mass of snow in a small area will cause a slower melt than if it were distributed throughout the streets.

Waste water input and water extraction

Human intervention in the hydrological regime of a river may be in the form of extraction (for irrigation or potable supply) or additional water from waste water treatment plants. The amount of water discharged from a sewage treatment works into a river may cause a significant alteration to the flow regime. At periods of low flow, 44 per cent of the river Trent (a major river draining eastern England) may comprise water derived from waste water effluent (Farrimond 1980, quoted in Newson 1995). There are times when the river Lea (a tributary of the Thames, flowing through north-east London) is composed of completely recycled water, which may have been through more than one sewage works. Jones has a startling diagram (1997: 227, Figure 7.9) showing very large diurnal variations in the river Tame that can be attributed to sewage effluent flows from the city of Birmingham, England. In this case the lowest flow occurs at 6 a.m. with a rapid rise in effluent flow that by midmorning has boosted flow in the river Tame by around 40 per cent (≈2.3 cumecs) (Jones 1997).

The extra flow that a river derives from sewage effluent may be especially significant if the waste water effluent has been abstracted from another catchment. The water in the river Tame naturally flows into the river Trent before flowing into the North Sea on the east coast of England. A large amount of abstracted water for Birmingham comes from the Elan valley in Wales, a natural tributary of the river Wye which drains into the Bristol Channel on the west coast of England. So in addition to causing diurnal fluctuations in the river Tame downstream of Birmingham, the waste water effluent is part of a water transfer across Great Britain. In a study into low flows in the United Kingdom (i.e. England, Wales, Scotland and Northern Ireland), Gustard et al. (1992) identified that 37 per cent of flow gauges were measuring flow regimes subject to artificial influence such as abstraction, effluent discharge or reservoir regulation of the river. This degree of flow alteration is a reflection of the high degree of urbanisation and the high percentage of the population living in an urban environment in the United Kingdom.

ESSAY QUESTIONS

1. **Discuss how well the principles of Integrated Water Resource Management are applied to the management of a catchment near you.**
2. **Explain the way that human-induced climate change may affect the hydrological regime for a region.**
3. **Assess the role of land use change as a major variable in forcing change in the hydrological regime for a region near you.**
4. **Compare and contrast the impact of urbanisation to the impact of land use change on general hydrology within the country where you live.**
5. **Discuss the major issues facing water resource managers over the next 50 years in a specified geographical region.**

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Chapters by different authors looking at change in the UK.

Global Water Partnership (2004) *Catalyzing change: A handbook for developing integrated water resources management (IWRM) and water efficiency strategies*. Published by GWP.
A handbook on practical steps to achieving IWRM.

Holden, J. (ed.) (2014) *Water resources: An integrated approach*. Routledge, Abingdon.
An excellent collection of chapters by various authors on contemporary themes and challenges in water management.

Intergovernmental Panel on Climate Change (IPCC) (2014) *Climate change 2014*. Cambridge University Press, Cambridge.
Various reports are published by the IPCC, summaries of which can be found at www.ipcc.ch.

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An excellent text covering thematic issues in water resource management.

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