Global Water Cycle

Advanced article

Nigel W Arnell, *University of Southampton, UK*

Article Contents

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The global water cycle describes the circulation of water around the world from one store to another via repeated evaporation of water from land and the oceans and precipitation back to the surface. The greatest amount of water is stored in the sea, and most fresh water in ice sheets and groundwater. Much groundwater is decoupled from the current hydrological cycle, having been formed in earlier wetter periods. The 'mobile' part of the cycle comprises only a small proportion of the total water.

Introduction

The global water (hydrological) cycle describes the circulation of water around the world from one store to another. This cycle consists of the repeated evaporation of water from land and the oceans and its precipitation back to the surface. River runoff and groundwater flow move precipitated water from one location to another, and atmospheric moisture transport moves evaporated water away from its source. The cycling of water and its flow across and under the land surface is central to weather and climate, and also to terrestrial and coastal zone ecosystems. Water is also vitally important to human life, society and economy, and over the centuries there have been many interventions in the hydrological cycle in response to both deficit and surplus of water. These interventions, together with changes in the vegetation cover within a catchment, have in many catchments had significant impacts upon the hydrological cycle.

This article describes the components of the global water balance, focusing on processes operating over the land surface. It reviews the effects of land surface characteristics on evaporation, and runoff generation processes, before describing some of the main hydrological regimes. The impacts of various human interventions on the water balance are explored, and a review of current and future pressures on water resources concludes. The article concentrates on the quantity of water. It is important to emphasize that rivers transport material from land to sea and that many key biochemical transformations take place in water: the hydrological system is a central component of many biogeochemical cycles. Human activities can substantially affect water quality, and quality is a major determinant of the suitability of a water source for many uses.

Components of the Water Balance

Global stores and fluxes

Figure 1 shows an estimate of the average annual fluxes of precipitation and evaporation over land and sea, together with runoff and atmospheric moisture transport (in units of $10^3\,km^3\,a^{-1}$) (Berner and Berner, 1987). There is, however, considerable variability in these fluxes over both space and time, and **Figure 2** shows the spatial distribution of average annual precipitation, evaporation and runoff and the ratio of runoff to precipitation.

Water moving through the hydrological cycle passes through a number of different stores, and estimates of the magnitudes of these stores are shown in **Table 1**. By far the greatest amount of water is stored in the sea, and most of the fresh water is stored in ice sheets (primarily in Antarctica and Greenland) and as groundwater. Estimates for the amount of water stored in groundwater are highly uncertain, although most are at the lower end of the range shown in **Table 1**. Much of this groundwater is held in long-term store, and indeed much is decoupled from the current hydrological cycle, having been formed in earlier, wetter periods. The 'mobile' part of the hydrological cycle therefore comprises a small proportion of the water on the planet.

Hydrological processes in the catchment

A catchment is defined as the area of land from which precipitation drains into a river (also known as a watershed

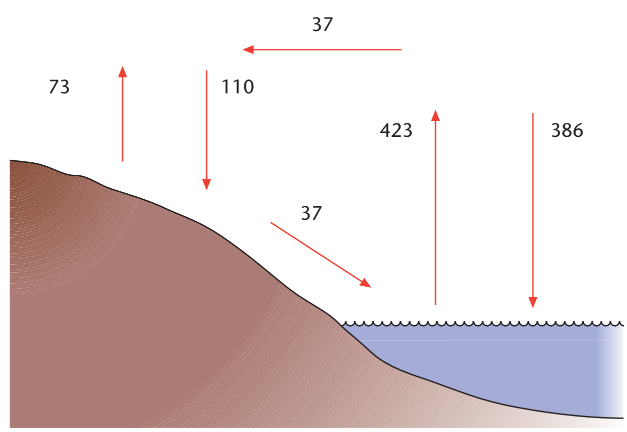

Figure 1 Fluxes of water in the global water cycle ($10^3\,km^3 a^{-1}$). Data from Berner and Berner (1987).

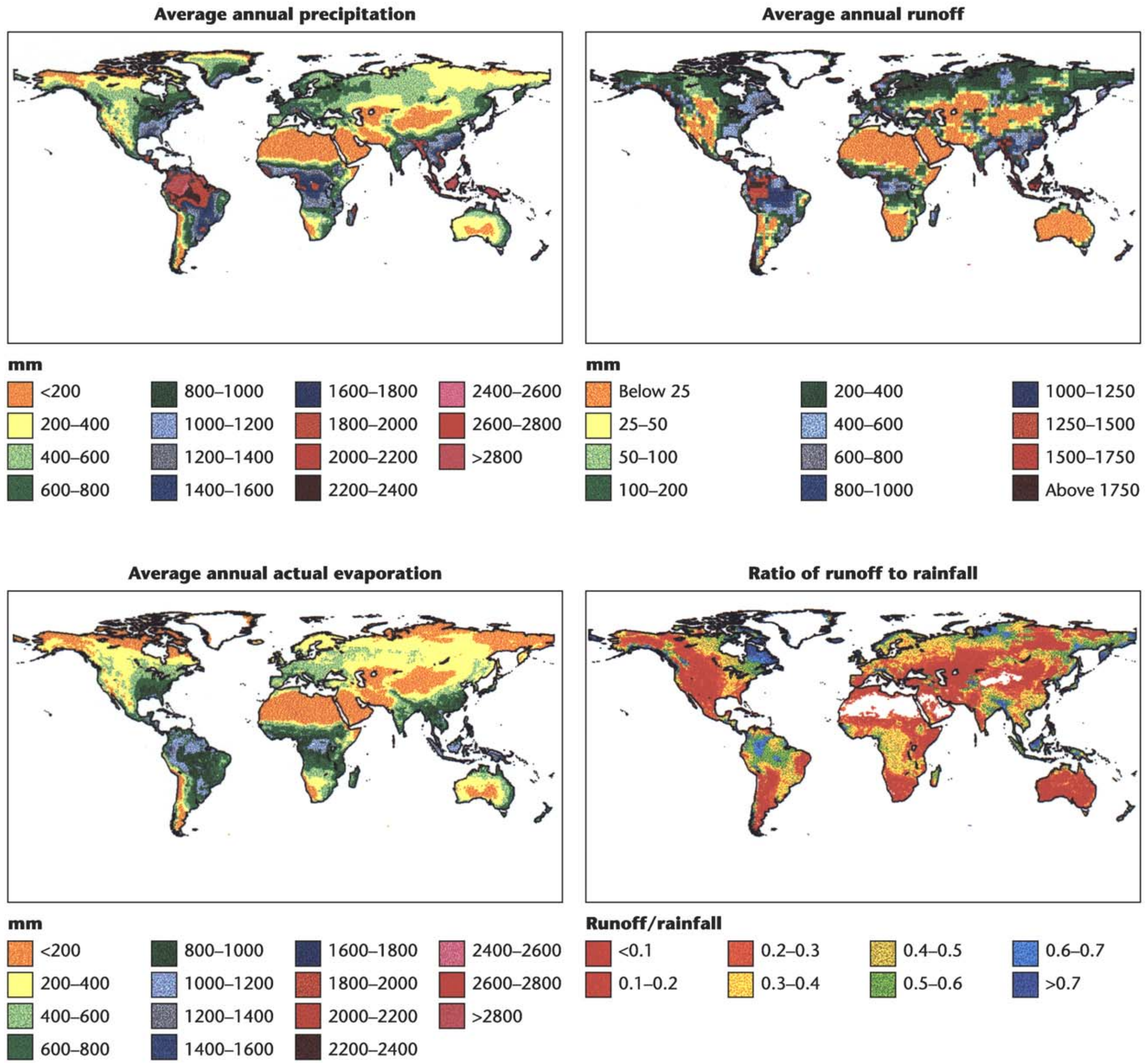

Figure 2 Distribution of average annual precipitation, evaporation and runoff, and the ratio of runoff to rainfall. Precipitation data from New *et al.* (1999), runoff from Korzun (1978).

in American usage; note that in British usage the watershed is the divide between two catchments); a river basin is simply a 'large' catchment. Catchments are usually defined on the basis of topography, although there are two complications. First, groundwater catchment limits are based on the height of the water table, which may not follow the surface topography. The effective area of a catchment underlain by chalk, a very important aquifer, is often very different from the apparent topographic catchment. Second, catchments may contain pockets of internal drainage, into which water drains but does not drain out. This is most common in areas of low relief.

Figure 3 shows the hydrological processes operating within a catchment. The relative importance of these processes varies over time and from region to region.

Precipitation

Precipitation is the condensation of water vapour into liquid water droplets and ice particles that fall to the Earth's surface. Rain is precipitation that is formed by water vapour condensing at temperatures above freezing or by ice crystals thawing before they reach the surface; snow falls when ice crystals do not thaw. In hydrological terms the

Table 1 Estimated storage of water

Store	Volume (10^3 km^3)
Oceans	1 320 000–1 370 000
Atmosphere	13–14
Land	
Ice caps and glaciers	24 000–29 000
Groundwater	8000–60 000
Freshwater lakes	91–125
Saline inland seas	85–104
Soil water	16.5–85
Water stored in permafrost	300
Rivers	1.2–2.1
Swamps and marshes	11.5
Biological water	0.6–1.1
Total land	36 000–84 400

Source: Data tabulated by Gleick (1993).

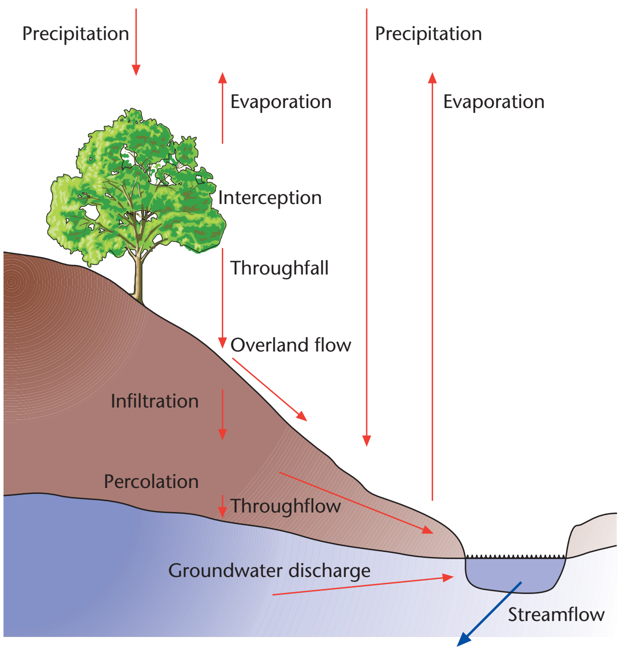

Figure 3 The hydrological cycle in a catchment.

distinction between rain and snow is important because snow is stored on the ground after it falls and the water is released only when the snow melts: this can be several months after the snow has fallen.

Details of condensation processes can be found in standard meteorological texts (e.g. Barry and Chorley, 1998), and it is important to note here only that precipitation forms when air containing moisture is cooled. There are three major mechanisms for cooling, each producing a distinctive pattern of precipitation over space and time. When air masses converge, the warmer air mass rises over the denser, cooler air mass, and cools. This is characteristic of mid-latitudes, creating 'depressions' that cover several thousand square kilometres, have a lifespan of 4–7 days, and move with the predominant air flow. They bring prolonged, widespread but relatively low-intensity precipitation. Nonfrontal convergence occurs within the same air mass, but where two wind systems converge: the Inter-Tropical Convergence Zone (ITCZ) is the boundary between northern and southern hemisphere wind systems, and migrates through the year with the thermal equator, bringing widespread rainfall. The second cooling mechanism is convection, when heating at the Earth's surface forces air to rise. Convective uplift is localized, and tends to produce short-duration, high-intensity rainfall. It is characteristic of areas with stable weather patterns and strong heating. Orographic uplift and cooling occur when an air mass is forced to rise by a topographic obstacle, such as a range of mountains.

Most precipitation over an area derives from moisture advected in from upwind. In some regions and in some seasons, however, a relatively large proportion is generated from 'local' evaporation, with the amount depending on the definition of 'local' (Trenberth, 1998). Approximately a quarter of the precipitation within the Amazon basin derives from evaporation within the basin, for example, and almost a third of Sahelian rainfall is locally generated in some seasons (Brubaker *et al*., 1993).

Interception and evaporation

A proportion of incoming precipitation is intercepted by vegetation before it reaches the surface. Some of this falls off the canopy as 'throughfall' and a small proportion runs down the stem of the vegetation as 'stemflow'. The rest of the intercepted water is evaporated (**Table 2**). Vegetation type, particularly its leaf area and branch density, has an important influence on the amount of throughfall and therefore of water reaching the surface, but so does precipitation type. The more intense the rainfall, the greater the proportion that reaches the surface as throughfall; snow is more easily intercepted than rain. Throughfall from one layer of the canopy may be intercepted by another lower down, and in many forest environments the litter layer or ground moss can intercept water and prevent it reaching the soil.

Evaporation is the conversion of a liquid to a vapour, and requires energy. This energy is largely provided by net radiation at the surface. The sun provides incoming short-wave radiation, some of which is reflected back to space by clouds and the atmosphere. A proportion of the incoming short-wave radiation that reaches the surface is reflected back to the atmosphere, depending on the reflectivity of the surface (as measured by 'albedo'). This reflectivity depends on surface colour, which depends on soil type, soil moisture content (wet soil is darker) and vegetation. The surface of

Table 2 Throughfall, stemflow and evaporation of intercepted water

Vegetation type	Location	Percentage of rainfall: Throughfall	Stemflow	Evaporation	Reference
Scots Pine	Norfolk, England	67	2	31	Gash and Stewart (1977)
Beech	Hampshire, England	82	4	14	Neal *et al*. (1993)
Eucalyptus	Southeast Australia	85	4	11	Crockford and Richardson (1990)
Cloud forest	Puerto Rico	59	2	39	Scatena (1990)
Tropical evergreen	Indonesia	87	1	11	Asdak *et al*. (1998)

the Earth emits long-wave radiation, some of which is reflected back to the surface from the lower atmosphere. Net long-wave radiation is generally negative and is relatively consistent through the year; net short-wave radiation varies with sun angle. An additional source of energy for evaporation may be heat energy advected in from upwind.

Evaporation will only occur if the air overlying the evaporating surface is not saturated. The amount of water vapour that air can hold increases with temperature, and is measured as the saturation vapour pressure (SVP). The ratio of the actual vapour pressure to the SVP is the relative humidity, and the difference between the two is the saturation vapour pressure deficit. Other things being equal, the greater the saturation vapour pressure deficit, the greater the rate of evaporation. Still air soon becomes saturated, but the rate of removal of air from the evaporating surface increases with windspeed and turbulence.

Energy, humidity and turbulence are therefore the three key drivers for the rate of evaporation from a surface, and if there is no shortage of water these determine the 'potential evaporation'. The rate of evaporation from open water is influenced also by the amount of energy stored in the water; in large, deep lakes the peak evaporation may be in early autumn rather than at the time of peak energy inputs. Evaporation from bare soil is determined not only by energy, humidity and windspeed, but also by the amount of water available, particularly in the top few centimetres of the soil: the rate of 'actual evaporation' may therefore be substantially lower than the rate of potential evaporation, and may decline rapidly once soil moisture contents fall.

There are two components to the evaporation from a vegetated surface. The first is the evaporation of intercepted water. This is determined by the amount of water stored on the vegetation, but also by the available energy, vapour pressure deficit and turbulence. The vegetation cover affects energy through its effect on albedo, and can also trap heat energy, which can mean that evaporation of intercepted water can occur during the night. The vegetation cover also influences turbulence and the rate of removal of water from the interception store. The second component of evaporation from a vegetated surface is transpiration, the evaporation of water through stomata in the leaves as part of the gas exchange process. The transpiration stream brings water to the plant that is used in photosynthesis, to maintain turgidity and bring dissolved minerals from the soil; the rate of transpiration is largely determined by the meteorological controls on evaporation. However, vegetation can exert an important influence. Root density obviously affects the amount of water that can be accessed by a plant to sustain transpiration through dry periods. The ability of stomata to allow transpiration (stomatal conductance) varies between plant types and, for a given plant type, can vary with five key factors: light intensity, CO_2 concentration, the leaf–air vapour pressure difference, leaf temperature, and leaf water content. The first two are relatively unimportant. As long as it is daylight, stomatal conductance is little affected by light intensity, and CO_2 concentration does not vary over short time scales. The other three factors are more important over the short term, and stomatal conductance tends to decrease as water stress increases.

The total evaporation from an area is made up of evaporation from open water, bare soil and intercepted water, together with transpiration. The relative importance of each component varies with land cover mix and also over time. In upland Britain, for example, frequent rain means that interception stores are often full, and the total evaporation of intercepted water is higher over a year than transpiration. In lowland Britain, however, less frequent rain means that the same vegetation type is wet less often, and annual transpiration tends to be greater than the annual evaporation of intercepted water. Where the vegetation is sparse, evaporation from soil may represent a large proportion of areal evaporation, particularly when the soil is wet. In the Sahel, for example, evaporation from wet bare soil is much greater than transpiration from adjacent plants (Gash *et al*., 1997): immediately after rain, bare soil evaporation could account for over 50% of areal evaporation, but after a couple of days had fallen to less than 20%. Plant transpiration remains relatively constant.

It is difficult to measure evaporation directly in the field. During the 1990s, however, a number of high-precision micrometeorological instruments were developed that essentially measure evaporation by accurately measuring short-duration fluxes of energy and water vapour. These instruments have helped significantly in the understanding of the dynamics of evaporation and transpiration. Theoretical analysis of the controls on evaporation and transpiration have led to the development of predictive equations, particularly the Penman and Penman–Monteith equations (Shuttleworth, 1993), which are widely used to model evaporation rates.

Infiltration and streamflow generation

Water that reaches the catchment surface either runs off the surface or infiltrates into the soil, depending on the 'infiltration capacity' of the soil and the intensity of the rainfall. Infiltration capacity is a function of soil texture, soil water content and structural features such as the presence of cracks. Water that infiltrates the soil can evaporate, percolate vertically downwards to groundwater, or flow laterally through the soil towards a river channel. Water is held in soils against gravity by 'matric forces', which include capillary forces and electrostatic adsorptive forces. As a general rule, the smaller the pore size the more tightly the water is held within the soil, and the lower the water content the more tightly that water is held. The rate of movement of water through soil is dependent on the gradient in the forces holding water in soil, and is therefore a function of soil texture and soil water content. In a given soil, the rate of movement falls by several orders of magnitude as soil water content reduces from saturation. Water can also move through cracks and voids in the soil, as macropore flow. This can be very rapid.

Flow in a river is therefore made up of overland flow that has run across the surface of the catchment, 'throughflow' that has flowed through the soil, and groundwater discharge, together with a very small amount of precipitation that falls directly into the river channel. The relative importance of these pathways varies between catchments and over time. A useful distinction is between 'quickflow' (water that reaches the river quickly) and 'baseflow' (water that sustains river flows during dry periods). Baseflow is largely derived from groundwater discharge and slow drainage of soil water.

In temperate environments, infiltration capacities are generally high, so rain water or melting snow will infiltrate into the soil. However, when the top of the soil becomes saturated, 'saturation-excess overland flow' will be generated, and water will move across the surface of the catchment and quickly reach a river channel. As the saturated area expands, a greater proportion of the catchment will generate saturation-excess overland flow, the channel network will expand upslope, and water will begin to move rapidly through the saturated soil. The rate of response of streamflow to an input of precipitation therefore depends on the extent of saturation in a catchment, and the rate at which these saturated areas expand. The most important implication of this is that a given input of precipitation has different effects depending on the state of the catchment. Studies with tracers have shown that most of the water in the initial phases of streamflow response to rainfall is not 'new' water but old water that has been pushed out of the soil by the input of new rain. As time progresses, an increasing proportion of the streamflow hydrograph is made up of new water.

Overland flow caused by rainfall exceeding soil infiltration capacity is also rare in humid tropical environments, because infiltration capacities tend to be very high. Response to rainfall is therefore, as in temperate areas, largely generated by flow through saturated soils and saturation-excess overland flow, with the major difference that wet contributing areas are more widely distributed through a humid tropical catchment because conditions are generally wetter.

The high intensity of rainfall and the thin soils in semi-arid areas together generate 'infiltration-excess overland flow', particularly where the soil develops a crust on being wetted; this overland flow may be very localized. However, not all this overland flow will reach a river, because downslope it may infiltrate into soils with higher infiltration capacity – perhaps soils that were not wetted by rainfall. Also, once water reaches a river channel in a semi-arid area, it may infiltrate into the river bed such that flows decrease downstream. This transmission loss recharges shallow aquifers, which may be depleted by evaporation. A given patch of semi-arid landscape will generate the same amount of overland flow for the same amount of rainfall, because antecedent conditions are not important. However, the amount of overland flow and river runoff generated from a rain storm depends on where the rain falls in the catchment, and particularly on the infiltration capacity of the soil on which the rain falls. Runoff generation in arid areas is similarly largely caused by infiltration-excess overland flow, but little may reach a river channel.

Runoff generation in 'cold' environments is characterized by the role of snow or ice. Where precipitation falls as snow over winter, this water is only released as the snow melts in spring. The melting water may infiltrate into the soil and either recharge groundwater or flow towards a stream as throughflow or, if the surface is impermeable, as runoff directly towards the river channel. The surface may be impermeable because it froze before the snow fell. In arctic and sub-arctic catchments the presence of permafrost affects runoff generation processes. The melting of the active layer releases water, and the presence of ice affects flow pathways and provides a barrier to percolation.

River flow regimes

It is clear from the above that different processes operate in different climatic and geological environments. These different combinations of processes produce different patterns of river runoff over time.

At the largest scale, the river flow regime is a function of climate. **Figure 4** shows average monthly runoff for four example catchments. In the temperate catchment, river flows are at a peak during winter. This is because although rainfall is evenly distributed through the year, evaporative demands are highest during summer and little of the summer rainfall generates runoff. The continental example shows the effect of snowfall and snowmelt: there is little runoff during winter, when precipitation falls as snow, but there is a large peak in spring. In the semi-arid example, streamflow is essentially generated during the wet season, and the timing of runoff therefore depends on the timing and duration of the wet season. In the tropical example, flows are high throughout the year because rain is frequent.

At the more detailed level, the variation in streamflow in a river from day to day is a function not only of catchment climate (such as the frequency and intensity of rainfall), but also catchment geology. **Figure 5** shows daily flows in one year for two adjacent catchments in southern England. One catchment is highly sensitive to rainfall, and has many short-duration streamflow peaks. The other catchment shows little response to the rainfall, and this is because it is underlain by very permeable chalk: virtually all rainfall infiltrates and percolates to groundwater, and very limited areas of saturation can be built up. Streamflow is therefore generated by the slow drainage of groundwater.

While it has long been recognized that hydrological regimes have varied over geological time scales, hydrologists and water resources managers have conventionally assumed that each year can be assumed to be a sample from a stationary population. However, it is increasingly recognized that persistent patterns of climatic anomalies – such as those associated with the El Niño Southern Oscillation phenomenon in the Pacific and Indian Oceans, the Pacific Decadal Oscillation, and the North Atlantic Oscillation – translate into persistent hydrological anomalies in many parts of the world, and that hydrological behaviour in many regions is closely correlated. During El Niño events, for example, streamflow tends to be higher than average in western South America, southeast South America, the Pacific southwest of the United States, north central United States and the Gulf Coast of the United States. Streamflow tends to be lower than average in Australasia, central and northern South America (including the Amazon), the Pacific northwest of North America, the northeast of North America, and equatorial and southern Africa.

Human Impacts on the Water Balance

Introduction: types of interventions

Humans have been affecting hydrological systems, directly and indirectly, for thousands of years, although in recent

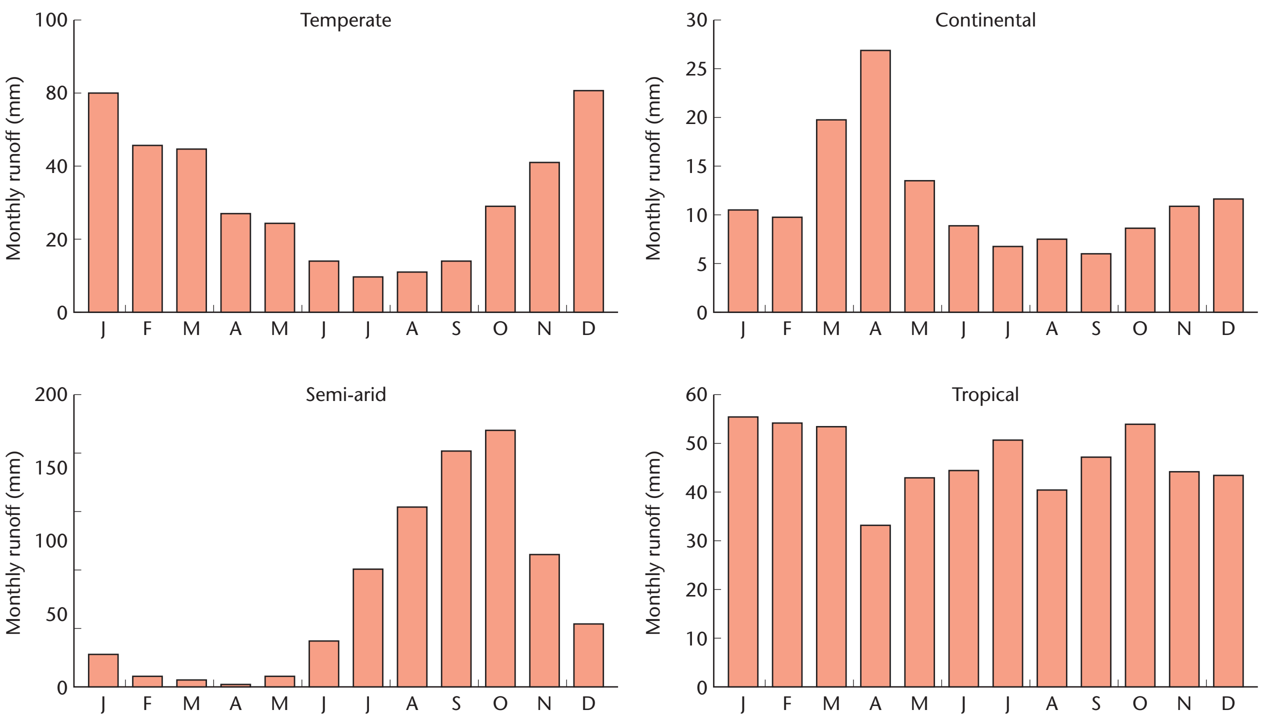

Figure 4 Four monthly runoff regimes.

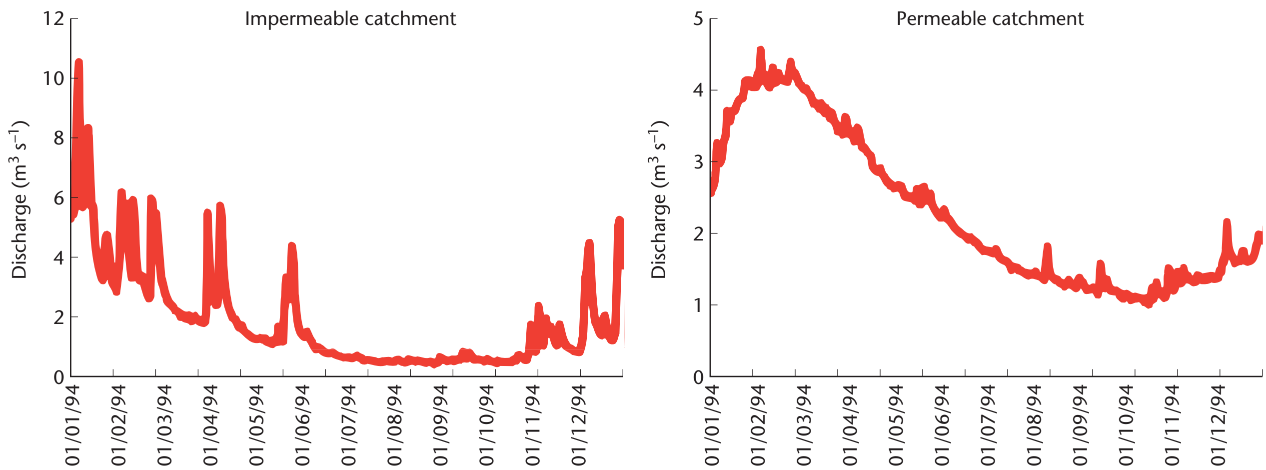

Figure 5 Daily flows for two British catchments.

decades the pace of change has accelerated. It is possible to distinguish between three different types of intervention: changes in land cover, abstraction and return of water, and physical infrastructure along the river network. To these changes 'on the ground' can be added global-scale changes in the inputs to a catchment through, for example, acid rain (which primarily affects water quality) and global warming. It is useful to distinguish between deliberate interventions aimed at changing hydrological behaviour and inadvertent effects resulting from some other action: deliberate interventions can also have inadvertent effects.

Changes in catchment land cover

The four key changes in land cover that have affected hydrological behaviour have been deforestation, afforestation, urbanization and agriculture. In general terms, these have the potential to affect the volume of flow and the timing of flow, through changes in the water balance and changes in runoff generation processes.

The primary hydrological effect of deforestation is to increase total runoff. This occurs because interception capacity is reduced and transpiration is reduced (particularly if the vegetation was deep-rooted and continued to transpire through a dry season). Flood peaks tend to be increased, with the greatest effects on the smaller floods: the frequency of large events tends to be relatively unaffected, because such events occur when the catchment is entirely saturated and land cover is not important. Removal of forest cover reduces the amount of rainfall necessary to generate quickflow, with the effect depending on the type of cover removed. The method of deforestation has been shown to have an effect on the extent of changes in runoff. Malmer (1996), for example, found that runoff in a Malaysian tropical forest catchment increased by around 20% if trees were removed using tractors and the land was cleared by burning, but increased by only 8% if the trees were removed manually and the land was not burnt. The removal of trees affected the water balance; the method of removal affected runoff generation processes, by compacting the soil and encouraging more overland flow. These hydrological changes are often associated with changes in erosion and sediment load. In large parts of Australia, deforestation has been associated with dryland salinization, caused by increased recharge and the subsequent rise in saline groundwater.

Although runoff as a whole tends to increase with deforestation, in some dry environments removal of forest cover is also associated with reductions in dry season flows. This is because wet season rainfall runs off rapidly and does not infiltrate to sustain flows through the dry season.

The effects of afforestation are in many senses the reverse of deforestation, and planting trees generally leads to a reduction in total runoff. However, the process of afforestation often involves the construction of roads and the draining of land, and these may in the short term lead to increased flood flows as water reaches the river channel more quickly.

Agriculture affects hydrological regimes partly through the abstraction of water for irrigation, but also by altering catchment land cover and encouraging physical changes to the landscape. The crop type planted affects the amount and duration of transpiration, and runoff generation processes may be altered by leaving land bare at certain times of the year. Important physical changes to the agricultural landscape that affect hydrological behaviour include the construction of roads and the installation of field drains. Roads generally enhance the response of the catchment to rainfall, but the effect of field drains is more ambiguous. In most cases, drains reduce the response to rainfall by increasing soil moisture deficits, but if the soil is saturated the drains serve merely to allow water to reach the river channel more rapidly.

The major effect of urbanization is to replace a largely permeable land surface with a largely impermeable one possessing a much larger (and permanent) drainage network. In practice, this makes a catchment behave much more like a low-infiltration-capacity, high-overland-flow semi-arid catchment: streamflow rises rapidly in response to rainfall, and also declines rapidly. As with deforestation, the largest floods are unaffected by urbanization because these occur when all the stores in a catchment are already full. Low flows in urbanized catchments are often sustained by the leakage of water from the supply and sewage pipes that underlie the catchment.

Abstraction and return of water

Water is taken from a river or aquifer ('abstracted') for municipal, industrial (including power generation) and agricultural use. This water may subsequently be returned to a river after use, although not necessarily the same river and usually not at the same quality. Virtually all water abstracted for municipal use is returned, but virtually none of the water abstracted for irrigation returns to the river as it is evaporated.

In principle, the hydrological effects of abstraction of water from rivers or reservoirs are relatively easy to assess, although in practice these hydrological changes and their ecological effects are often ignored by abstractors. Consequently, there are several examples of major rivers that have been heavily altered by cumulative abstractions. The two best examples are the Colorado River in the western United States, which in some years does not reach the sea in Mexico, and the rivers draining into the Aral Sea in central Asia – by the mid 1970 s, only between 10% and 20% of the average inflows were reaching the Aral Sea (Micklin, 1988), and the sea accordingly shrank.

The effects of abstraction of groundwater are rather more difficult to predict and assess. They depend on the rate of abstraction, the effect of the abstraction on groundwater levels, and the connectivity between the groundwater at the well-site and the river. In Britain, for example, cases of overabstraction from aquifers were only observed during the 1988–1992 drought, when the abstractions increased the effect of rainfall shortages in a number of high-profile cases. In a number of semi-arid areas – notably in parts of Spain – overabstraction of groundwater for irrigation is lowering water tables and reducing baseflows to rivers.

Physical infrastructure

Physical infrastructure that can alter both the volume and timing of flows includes reservoirs, weirs and diversion structures that transfer water from one river to another. Channel modifications, such as straightening or dredging, tend to affect only the timing of flows through a flood event.

It is estimated that there are around 45 000 'large' dams in the world storing around 6000 km^3 of water, and 46% of the world's 106 primary major river basins are modified by at least one dam (World Commission on Dams, 2000). The primary hydrological effect of a reservoir is to alter the timing of streamflow through the year, with the effect depending on the way the reservoir is operated. In general, reservoirs lessen flow variability downstream. Evaporation from reservoirs in dry regions can substantially reduce the volume of flow, especially when combined with abstractions for irrigation and other uses.

Changes to catchment inputs

Climate change due to an increasing concentration of greenhouse gases has the potential to affect hydrological regimes, depending partly on the actual change in precipitation and evaporative demand, and partly on catchment characteristics (Arnell, 1999; Arnell and Liu, 2001). Higher temperatures tend to increase evaporative demands largely through an increase in the vapour pressure deficit, but this may be either offset or exaggerated by changes in net radiation, humidity and windspeed. Increasing CO_2 concentrations and a change in rainfall and temperature will also impact on vegetation characteristics and hence evaporation, although these effects are currently uncertain and in many catchments will be outweighed by management decisions. There have been many catchment-scale assessments of the potential effects of climate change on river flows (and rather fewer on recharge), showing the sensitivity of estimated changes to assumed changes in rainfall. A rather more robust conclusion, however, is that increasing temperatures would lead in many areas to reductions in the amount of snowfall, and hence increased flows during winter and reduced flows in spring. Such a change in the timing of streamflow would have significant implications for many users of river flows, such as abstractors, power generators and navigators.

Impacts beyond the catchment

Hydrological processes affect the earth system by influencing fluxes of energy and water in the atmosphere and by determining fluxes of water to the coastal zone. It is increasingly recognized that human interventions in the hydrological system can have consequences beyond the catchment.

The amount of water on the land surface affects the atmosphere not only by providing a source of moisture for evaporation but also by altering the energy balance at the surface: the more energy used for evaporation, the less used for sensible heating. Changes in catchment land cover, if sufficiently extensive, have the potential to affect regional-scale energy and water fluxes and hence regional and local climate. For example, there is evidence that deforestation in parts of the Sahel exaggerated the effect of a lack of rainfall by reducing the amount of precipitation recycling (Nicholson, 2000), and Sud *et al.* (1996) have simulated how past deforestation of parts of the Amazon rainforest have led to an 8% reduction in basin-wide evaporation. Reduction in the extent of wetlands in semi-arid areas, either through drainage or upstream reservoirs, has the potential to reduce fluxes of water to the atmosphere, and thus alter local and regional climate.

River inflows to the coastal zone can have very significant influences on coastal processes of sedimentation and deposition; changing these inflows can alter these processes, and there are many documented cases of shoreline retreat associated with impoundment and its effects on both the volume of flow and the amount of sediment transported to the coast. A reduction of flows by around 28% in the Ebro River in Spain, for example, due to irrigation and evaporation from reservoirs, has led to an increase in both the penetration of the salt wedge into the estuary of the Ebro and the duration of saline conditions (Ibanez *et al.*, 1996). Changes in the volume of flows have also been associated with alterations in ocean circulation patterns in the Baltic (Schinke and Matthaus, 1998) and Mediterranean (Turley, 1999) seas. In the latter case, lower inflows from the Nile have led to increased incursions of relatively fresh water from the Atlantic, and increased outflows of saline water.

At the global level, Gornitz *et al.* (1997) estimated that the net effect of actions that increased runoff (such as deforestation) and decreased runoff (such as evaporation from reservoirs and irrigation) was to have lowered sea level by around 1 cm during the twentieth century, with most of the decrease in the last few decades.

Human interventions: an overview

This section has reviewed in general terms some of the potential impacts of human activities on river flows, and their consequences beyond the catchment. These changes are superimposed on to natural climatic variability, and an apparent increase in frequency of 'unusual' events is not necessarily a sign of increasing human impact. These interventions may counteract or reinforce each other. Finally, however, it is important to emphasize that many management strategies are now in place in many catchments to attempt to mitigate the worst effects of human interventions in the hydrological system.

Human Use of Water Resources

The previous section introduced the concept of abstractions and returns of water. There are, in fact, many human uses of the water environment, which are often classified as being 'offstream' or 'instream'. Offstream uses take water away from the river (or aquifer), and include abstractions for public and municipal supply, industrial usage and power generation, and irrigation. This water is sometimes returned to the river. Instream uses exploit the water within the river channel, and include hydropower generation, navigation, recreation and, perhaps most generally, waste disposal. **Figure 6** shows the trend in global water abstractions over the last 50 years, showing that agriculture is by far the dominant user. However, **Figure 6** hides very significant variation between regions. In some areas – such as the United Kingdom – agriculture abstracts very little, and in other areas industrial demands for cooling are dominant. Also, in parts of Europe and North America, total demand for water is falling, as abstractors use water more efficiently.

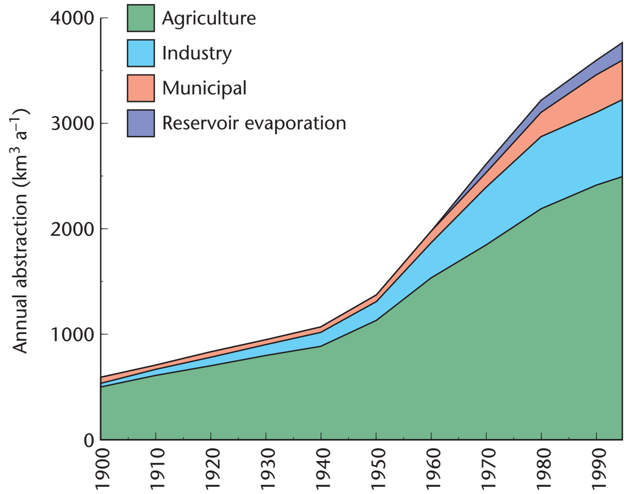

Figure 6 Global abstraction of water. Data from Shiklomanov (1998).

Water management faces a number of challenges over the next few decades. Many of these require technical solutions but most require institutional change and alterations to the way in which water is managed. There are three main concerns. In many countries there is a need to ensure that supplies continue to be available to meet demands. Conventionally, this has taken the form of measures to increase supply, but increasingly water management organizations are attempting to curb demand. This is partly for cost reasons, but largely because of environmental concerns, and a shift towards the management of the water environment rather than the needs of human users is occurring in many countries. The second challenge for water management is to provide safe, clean water for all, which focuses attention not only on water supply but also on effluent treatment, the distribution of water and, very importantly, access to water. The third challenge is to prevent and reverse the degradation of the water environment by overabstraction or pollution. All these challenges need to be met in the face of an increasingly uncertain future – economic, political and hydrological – and must be based on a sound scientific understanding of hydrological processes.

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