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1 Introduction

1.1 Definition and Scope

Hydrology is literally the science of water. Etymologically, the word has its roots in ancient Greek, and is a composite, made up of *ὕδωρ*, water, and λόγος, word. Obviously, defined this way, the term is much too broad to be very useful, as it would have ramifications in all scientific disciplines.

Actually, the word hydrology has not always been well defined and even as recently as the 1960s it was not very clear exactly what hydrology was supposed to cover and encompass. Price and Heindl (1968), in a survey of many of the definitions that had appeared in the literature over the previous 100 years, were compelled to conclude that the question "What is hydrology?" had not been resolved by their review. Still, they felt that, in general, there seemed to be a consensus that hydrology is a physical science, which is concerned mainly with the water cycle of land and near-shore areas; moreover, there had been a tendency to broaden the term rather than to narrow it, even to the point of including socio-economic aspects.

Over the past few decades, however, with the growing activity level and the increasing maturity of this field of endeavor, a more precise definition has emerged. Hydrology is now widely (see, for example, Eagleson, 1991) accepted to be the science that deals with those aspects of the cycling of water in the natural environment that relate specifically to:

– *the continental water processes*, namely the physical and chemical processes along the various pathways of continental water (solid, liquid, and vapor) at all scales, including those biological processes that influence this water cycle directly; and to
– *the global water balance*, namely the spatial and temporal features of the water transfers (solid, liquid, and vapor) between all compartments of the global system, i.e. atmosphere, oceans, and continents, in addition to stored water quantities and residence times in these compartments.

Because it is defined as being concerned specifically with continental water processes, hydrology is a discipline distinct from meteorology, climatology, oceanology, glaciology, and others that also deal with the water cycle in their own specific domains, namely the atmosphere, the oceans, the ice masses, etc., of the Earth; at the same time, however, hydrology integrates and links these other geosciences, in that, through the global water balance, it is also concerned with the exchanges of water between all these separate compartments.

With this definition it is now also possible to delineate the practical scope of hydrologic analysis in engineering and in other applied disciplines. It consists of the determination of the amount and/or flow rate of water that will be found at a given location and at a given time under natural conditions, without direct human control or intervention. The latter specification, that no human control be involved, is necessary to distinguish hydrology from the related discipline of hydraulics. Hydraulics is concerned with the study of controlled fluid motion in well-defined and often in human-made environments. For instance, problems involving pipe flow, irrigation water distribution, or pumping of groundwater are not hydrologic in nature, but are more properly assigned to the realm of hydraulics.

1.2 The Hydrologic Cycle

The water cycle, also called the hydrologic cycle, refers to the pathway of water in nature, as it moves in its different phases through the atmosphere, down over and through the land, to the ocean, and back up to the atmosphere. When atmospheric water vapor condenses and precipitates over land, initially it moistens the surface and some amount of it is stored as *interception*, which later evaporates. As *precipitation* (and in a similar way *snowmelt*) continues, part of it may flow over the surface in the form of *overland flow* or *surface runoff*, and part of it may enter into the soil as *infiltration*. This surface runoff soon tends to collect locally, either in puddles or small ponds as *depression storage*, or in gullies or larger channels where it continues as *streamflow*, which ultimately ends up in a larger water body, such as a lake or the ocean. Streamflow is normally described by a *hydrograph*, that is the rate of flow at a gaging station as a function of time. The infiltrated water may flow rapidly through the near-surface soil layers to exit into springs or adjacent streams, or it may percolate more slowly through the profile to join the *groundwater*, which sooner or later seeps out into the natural river system, lakes and other open-water bodies; part of the infiltrated water is retained in the soil profile by capillarity and other factors, where it is available for uptake by the roots of vegetation.

Soil layers and other geologic formations, whose pores and interstices can transmit water, are called *aquifers*. When an aquifer is in direct contact with the land surface, it is referred to as *unconfined*. The locus of points in an unconfined aquifer, where the water pressure is atmospheric, is called the *water table*. Although the water table is not a true free surface separating a saturated zone from a dry zone, to simplify the analysis it is sometimes assumed to be the upper boundary of the groundwater in an unconfined aquifer. The partly saturated zone in an unconfined aquifer, between the water table and the ground surface, is sometimes referred to as the *vadose zone*. In an unconfined aquifer, the term *groundwater* refers usually to the water found below the water table; *soil water* or *soil moisture* refers to the water above the water table. A water-bearing geologic formation, that is separated from the surface by an impermeable layer, is referred to as a *confined aquifer*. Streamflow is fed both by surface runoff and by subsurface flow from riparian (i.e. located along the banks) aquifers. The streamflow, resulting from groundwater outflow is often called *base flow*; in the absence of *storm flow* or *storm runoff* caused by precipitation, base flow is also referred to as *drought flow* or *fair-weather flow*.

Finally, the hydrologic cycle is closed by *evaporation*, which returns the water, while in transit in the different flow paths and stages of storage along the way, back into the atmosphere. When evaporation takes place through the stomates of vegetation, it can be referred to as *transpiration*. Direct evaporation from open water or soil surfaces and transpiration of biological water from plants are not easy to separate; therefore the combined process is sometimes called *evapotranspiration*. Although this term may on occasion be helpful and explanatory locally during the growing season in agriculture and forestry, it can be misleading for general usage as it fails to include such important vapor flux components as evaporation from intercepted water due to ample rainfall, from small open-water bodies scattered in the landscape, and from snow and ice during winter in many regions of the world. Evaporation of ice is referred to as *sublimation*. While these distinctions can be useful at times, the term *evaporation* is normally not only adequate but also

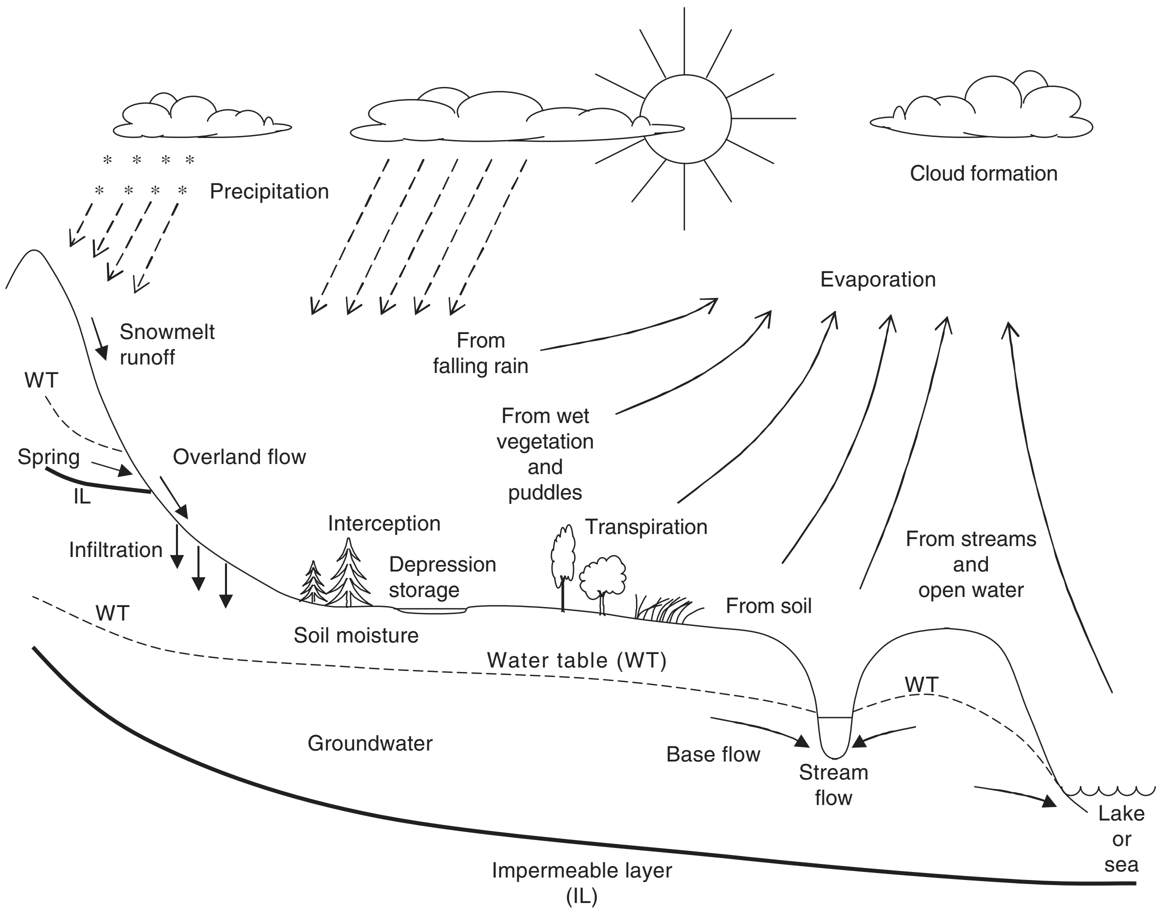

Figure 1.1 Sketch of some of the main processes in the land phase of the water cycle.

preferable to describe all processes of vaporization (Miralles et al., 2020). Some of the main processes are drawn schematically in Figure 1.1.

1.3 Some Estimates of the Global Water Balance

Numerous studies have been carried out to estimate the magnitude of the most important components of the water-budget equations on a global scale. Because the available database required for this purpose is still far from adequate, several of the methods used in these estimates may be open to criticism. Nevertheless, there is a fair agreement among some of the calculated values and, within certain limits, they provide a useful idea of the long-term average balance in different climatic regions of the world.

As shown in Table 1.1, for the entire Earth, the average annual precipitation and evaporation have the same magnitude, i.e. $P = E$, and are of the order of 1 m. However, their average values of P and E over the Earth's oceans and land surfaces are markedly different. Water is continuously removed from the ocean into the atmosphere by the small surplus of E over P, and this is counterbalanced by the larger surplus of P over E on land. Under steady conditions, that is for long time periods, the latter surplus, or

Table 1.1 Estimates of world water balance (m a^{-1})

	Land (1.49×10^8 km^2)			Oceans (3.61×10^8 km^2)		
Reference	P	R	E	P	E	Global $P = E$
Budyko (1970, 1974)	0.73	0.31	0.42	1.14	1.26	1.02
Lvovitch (1970)	0.73	0.26	0.47	1.14	1.24	1.02
Lvovitch (1973)	0.83	0.29	0.54	–	–	–
Baumgartner and Reichel (1975)	0.75	0.27	0.48	1.07	1.18	0.97
Korzun et al. (1978)	0.80	0.315	0.485	1.27	1.40	1.13
Rodell et al. (2015)	0.80	0.31	0.48	1.11	1.23	1.02

Table 1.2 Some estimates of the mean precipitation (and river runoff) for the continents (in m a^{-1})(*)

	Europe	Asia	Africa	North America	South America	Australia and Oceania	Antarctica
Percent of land area	6.7	29.6	20.0	16.2	12.0	6.0	9.5
Reference							
Lvovitch (1973)	0.734	0.726	0.686	0.670	1.648	0.736	–
	(0.319)	(0.293)	(0.139)	(0.287)	(0.583)	(0.226)	–
Baumgartner and Reichel (1975)	0.657	0.696	0.696	0.645	1.564	0.803	0.169
	(0.282)	(0.276)	(0.114)	(0.242)	(0.618)	(0.269)	(0.141)
Korzoun et al. (1977)	0.790	0.740	0.740	0.756	1.600	0.791	0.165
	(0.283)	(0.324)	(0.153)	(0.339)	(0.685)	(0.280)	(0.165)

*The corresponding evaporation values can be determined with Equation (1.1).

remainder, can be considered to be runoff from the Earth's rivers into the oceans, R (expressed as height of water column per unit of time). Thus, over the Earth's land surfaces the average annual water balance can be described by

$$R = P - E \tag{1.1}$$

Over the land surfaces the average precipitation intensity, P, is about 0.80 m a^{-1}, whereas the corresponding average evaporation, E, is around 0.50 m a^{-1} or about 60 to 65% of the precipitation. Averaged over all continents and over long time periods, the annual runoff, R, is therefore around 35 to 40% of the precipitation. Except for South America and Antarctica (see Table 1.2), the values for the individual continents are not very different from the global values. Precipitation and streamflow runoff measurements have been and are being made routinely in many places on Earth. In contrast, evaporation has not received as much attention and only in recent years have attempts been made to obtain systematic measurements (Pastorello et al., 2017).

Estimates of the average distribution of water in different forms expressed as depth of water covering the globe, assumed to be a perfect sphere, are given in Table 1.3. These indicate that the 1 m of average annual precipitation (shown in Table 1.1) is relatively large as compared to the active fresh water on Earth, that is the water which is not stored in permanent ice and deep groundwater. This means that the turnover

Table 1.3 Estimates of different forms of global water storage (as depth in m over entire Earth's surface)

Source of data	Lvovitch (1970)	Baumgartner and Reichel (1975)	Korzun et al. (1978)
Oceans	2686	2643	2624
Ice caps and glaciers	47.1	54.7	47.2
Total groundwater	117.6	15.73 (excluding Antarctica)	45.9
(Active groundwater)	(7.84)	(6.98)	–
Soil water	0.161	0.120	0.0323
Lakes	0.451	0.248	0.346
Rivers	0.002 35	0.002 12	0.004 16
Atmosphere	0.0274	0.0255	0.0253

of the active part of the hydrologic cycle is rather fast, and that the residence times in some of the major compartments of the water cycle are relatively short; the mean residence time can be taken as the ratio of the storage and the flux in or out of storage. For example, a continental runoff rate of 0.30 m a^{-1} (Table 1.1) and a storage in the rivers of (0.003/0.29) m of water on the 29% of the world occupied by land, gives a mean residence time of the order of 13 days for the rivers of the world. Similarly, a global evaporation rate of 1 m a^{-1}, with 0.025 m of storage in the atmosphere, leads to a mean residence time of the order of 9 days for the atmosphere. These are very short residence times. Moreover, as the oceans occupy about 71% of the Earth's surface, the active fresh water in the hydrologic cycle is continually being distilled anew through ocean evaporation.

Maps depicting the approximate distribution of components of the water balance in different parts of the world have been presented by, among others, Lvovitch (1973), Budyko (1974), Baumgartner and Reichel (1975), and Korzoun et al. (1977). The relative and absolute magnitudes of the main components of the hydrologic cycle, namely P, R, and E, can vary over a wide range from one location to another. Obviously, the long-term mean values of all three are negligible in desert locations. At the other extreme, maximal annual precipitation values of up to 26.5 m have been recorded in a mountainous monsoon environment (Cherrapunji, Meghalaya). Maximal mean evaporation values of up to 3.73 m a^{-1} have been inferred for the Gulf Stream in the western Atlantic (Bunker and Worthington, 1976) and up to 4 or even 5 m a^{-1} for the Gulf of Aqaba (Assaf and Kessler, 1976).

Much research has been directed in recent years into studying the evolution of today's climate in response to increasing greenhouse gases in the atmosphere. For some time now, there have been numerous indications of an accelerating hydrologic cycle in certain regions (Brutsaert and Parlange, 1998; Karl and Knight, 1998; Lins and Slack, 1999). But also globally, the fluxes shown in Table 1.1 are now known to be evolving in tandem with the gradual average rise in temperature of the atmosphere. For instance, an analysis of satellite measurements, mainly over the oceans supplemented with land data, by Wentz et al. (2007) has indicated that during the period 1987–2006 the average relative rate of increase of the total atmospheric water vapor W'/W, of the global precipitation P'/P, and of the global evaporation E'/E have been roughly the same, namely about 0.0013 a^{-1}. (Here the prime $(\ )'$ denotes the average time derivative $d(\ )/dt$.) During these two decades around the turn of the century, the global lower tropospheric warming rate was about $T' = 0.02\,^{\circ}\mathrm{C}\ \mathrm{a}^{-1}$, which means that the average trend of 0.13% per year amounts to about 6 to 7% per kelvin of warming, say 0.065 K^{-1}; this is nearly the same as the rate produced by the Clausius–Clapeyron (or C-C) equation (see Equation (2.13)), with constant relative humidity at a temperature of $14\,^{\circ}\mathrm{C}$,

Table 1.4 Estimates of mean global heat budget at the Earth's surface in W m^{-2}

	Land			Oceans			Global		
Reference	R_n	L_eE	H	R_n	L_eE	H	R_n	L_eE	H
Budyko (1974)	65	33	32	109	98	11	96	80	16
Baumgartner and Reichel (1975)	66	37	29	108	92	16	96	76	20
Korzun et al. (1978)	65	36	29	121	109	12	105	89	16
Ohmura (2005)	62	36	26	125	110	15	104	85	19
Trenberth et al. (2009)	65.5	38.5	27	110	97	12	98	80	17
Wild et al. (2015)	70	38	32	117	100	16	103	82	21

the commonly accepted global average temperature in the lower troposphere. Application of this C-C similarity in Brutsaert (2017) showed that the global rate of increase of the hydrologic cycle during the second half of the twentieth century was of the order of $P' = E' = 1.0$ mm a^{-2}; the evaporation trend from the global land surfaces was found to be about half this value, namely of the order of 0.4 to 0.5 mm a^{-2}. Very similar values were obtained in other studies by different methods (Brutsaert, 2006; Miralles et al., 2014; Anabalón and Sharma, 2017). In response to the global precipitation trend, the overall global runoff has been observed to increase as well. From the monthly discharges of the largest rivers worldwide, Labat et al. (2004) estimated this increase to be about 4% per kelvin; interestingly, this is roughly of the same order as the Clausius–Clapeyron value of 0.065 K^{-1} mentioned above for precipitation and evaporation. However, the runoff changes also showed large variability over the different continents, including negative trends in some regions (Zhang et al., 2015).

The strong link between the water cycle and climate is further illustrated by the estimates of the mean global surface energy budget in Table 1.4. Over large areas and over sufficiently long periods, when effects of unsteadiness, melt and thaw, photosynthesis and burning, and lateral advection can be neglected, this surface energy balance can be written as

$$R_n = L_e\, E + H \tag{1.2}$$

where R_n is the specific flux of net incoming radiation, L_e is the latent heat of vaporization, E is the rate of evaporation, and H is the specific flux of sensible heat into the atmosphere. The major portion of the incoming radiation is absorbed near the surface of the Earth, and is transformed into internal energy. The subsequent partition of this internal energy into longwave back radiation, upward conduction and convection of sensible heat, H, and latent heat, L_eE, is one of the main processes driving the atmosphere. Table 1.4 (see also Figure 2.28) indicates that the net radiative energy is mainly disposed of as evaporation. Over the oceans, the latent heat flux L_eE is on average larger than 90% of the net radiation. But even over the land surfaces of the Earth, L_eE is on average still larger than half of R_n.

Because the global patterns of heating force the circulation of the planetary atmosphere, the implications of this large latent heat flux are clear. As a result of the relatively large latent heat of vaporization, L_e, evaporation of water involves the transfer and redistributiuon of large amounts of energy under nearly isothermal conditions. Because, even at saturation, air can contain only relatively small amounts of water vapor, which can easily be condensed at higher levels, the air can readily be dried out; this release of energy through condensation and subsequent precipitation is the largest single heat source for the atmosphere. Thus processes in the water cycle play a central role in governing weather and climate.

1.4 Methodologies and Procedures

This book aims primarily to describe the occurrence and transport of water in its continuous circulation over the land surfaces of the Earth. Before starting this task, it is worthwhile to review briefly the different strategies that are available and that can be used for this purpose.

1.4.1 Statistical Analysis and Data Transformation

As observed in Section 1.2, one of the main practical objectives of hydrologic analysis is the determination of the quantity of water, in storage or in transit, to be found at a given time and place, free of any direct human control. When a reliable record of observed hydrologic data is available, a great deal can be learned simply by a statistical analysis of this record. Although such an approach is proper for stationary systems in the prediction of long-term behavior for general planning purposes, it cannot be used for short-term and emergency forecasting, for example, during floods, or for day-to-day resource-management decisions. Furthermore, reliable records are available for only a few locations over a limited period of time, and practically never where needed. Therefore, in hydrology the problem is often such that a method must be devised to transform some available data, which are of no direct interest, to the required hydrologic information. For instance, the problem may consist of determining the rate of flow in a river at a given location either from a known flow rate at some other point upstream or downstream, or from a known rainfall distribution over the upstream river basin. In other cases, the problem may consist of deducing the basin evaporation from soil and vegetation on the basis of available meteorological data.

1.4.2 The "Physical" versus the "Systems" Approach

The hydrologic literature is replete with attempts at classifying the methodologies and paradigms that have been used to transform hydrologic input into hydrologic output information. Until a few years ago it had become customary to consider two contrasting approaches, namely the "physical" approach and the "systems" approach. In the physical approach, the input–output relationship is sought by the solution of the known conservation equations of fluid mechanics and thermodynamics with appropriate boundary conditions to describe the flow and transport of water throughout the hydrologic cycle. This approach has obvious limitations; the physiographic and geomorphic characteristics of most hydrologic systems are so complicated and variable, and the degree of uncertainty in the boundary conditions so large, that solutions are feasible only for certain highly simplified situations. In other words, the properties of natural catchments can never be measured accurately enough, and solutions, based on internal descriptions starting from first principles of fluid mechanics, can be obtained only for grossly idealized conditions, which are coarse approximations of any real situation.

The hydrologic "systems" (also "operational" or "empirical") approach is presumably based on a diametrically opposite philosophy. In this approach the physical structure of the various components of the hydrologic cycle and their inner mechanisms are not considered; instead, each component, however it may be defined, is thought of as a "black box," and the analysis focuses on discovering a mathematical relationship between the external input (e.g. rainfall, air temperature, etc.) and the output (e.g. river flow, soil moisture, evaporation, etc.). The structure of this mathematical relationship is mostly quite remote from the physical structure of the prototype phenomena in nature. This lack of correspondence between the inner physical mechanisms and the postulated functional formalisms makes this approach quite general operationally, because it permits the use of well-known algorithms and objective criteria in identification and prediction. However, this also underlies the main limitations of this approach. First, in assigning cause and

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effect, the definition of input and output variables is mostly based on intuition guided by past experience, and the danger exists that some important phenomena are overlooked. Second, the best that can ever be expected with a black-box approach is a satisfactory reproduction of a previously obtained input–output record; even when such data are available, it is difficult to accommodate fully the nonstationary effects in the system, and it is impossible to anticipate subsequent hydrologic changes, such as those resulting from urbanization, deforestation, reclamation, or climate change.

Because many hydrologic methods do not really fit in this physical-versus-empirical classification, a third possible approach was taken to be an intermediate one. In this view the performance of a hydrologic unit, say a catchment, is represented in terms of some idealized components or "gray boxes," which correspond to recognizable elements in the prototype, whose input–output response functions are structured after solutions of some tractable or suitably simplified situations of the physical processes perceived to be relevant. This third way was often called the "conceptual model" approach.

At first sight, a classification based on three distinct approaches, namely physical, empirical, and conceptual, may appear reasonable. However, it is less than obvious how this classification can be applied to specific cases. Indeed, one might ask what the difference is between physical and empirical. After all, the essence of physical science is experimentation and conceptualization. Moreover, the physical approach of one discipline is usually the empiricism or the conceptual model of another. For example, Newton's "law" of viscous shear constitutes the physical basis of a wide area of fluid mechanics, whereas it represents a mere black-box simplification in molecular physics. Darcy's law is the physical basis of much of groundwater hydrology, but in fluid mechanics it can be considered an operational approach, to avoid the complexity of flow analysis in an irregular and ill-defined pore network. The same dilemma is inherent in most other special concepts used in hydrology. This ambiguous difference between physical, empirical, and conceptual shows that the classification of the methodologies should be based on other criteria.

1.4.3 Spatial Scale and Parameterization

General Approach

All natural flow phenomena are governed by the principles of conservation of mass, momentum, and energy, which can be expressed by a number of equations to provide a mathematical description of what goes on. However, because there are normally more dependent variables than available conservation equations, in order to close the system, additional relationships must be introduced. These closure relationships, also called *parameterizations*, relate some of the variables to each other to describe certain specific physical mechanisms; the mathematical form of these relationships, and the values of the material constants or *parameters* are usually based on experimentation.

A second point is that any physical phenomenon must be considered at a given scale; this scale is the available (depending on the data) or chosen (depending on the objectives of the study) resolution. It will become clear later on in this book that, while the fundamental conservation equations remain unaffected by the scale at which the phenomenon is being considered, most closure relationships within them are quite sensitive to scale. Indeed, a parameterization can be considered as a mathematical means of describing the subresolution (or microscale) processes of the phenomenon, in terms of resolvable scale (or macroscale) variables; these macroscale variables are the ones that can be treated explicitly in the analysis or for which measured records are obtainable. Thus, the details of the microscale mechanisms are not considered explicitly, but their statistical effect is formulated mathematically by a parameterization in terms of macroscale variables.

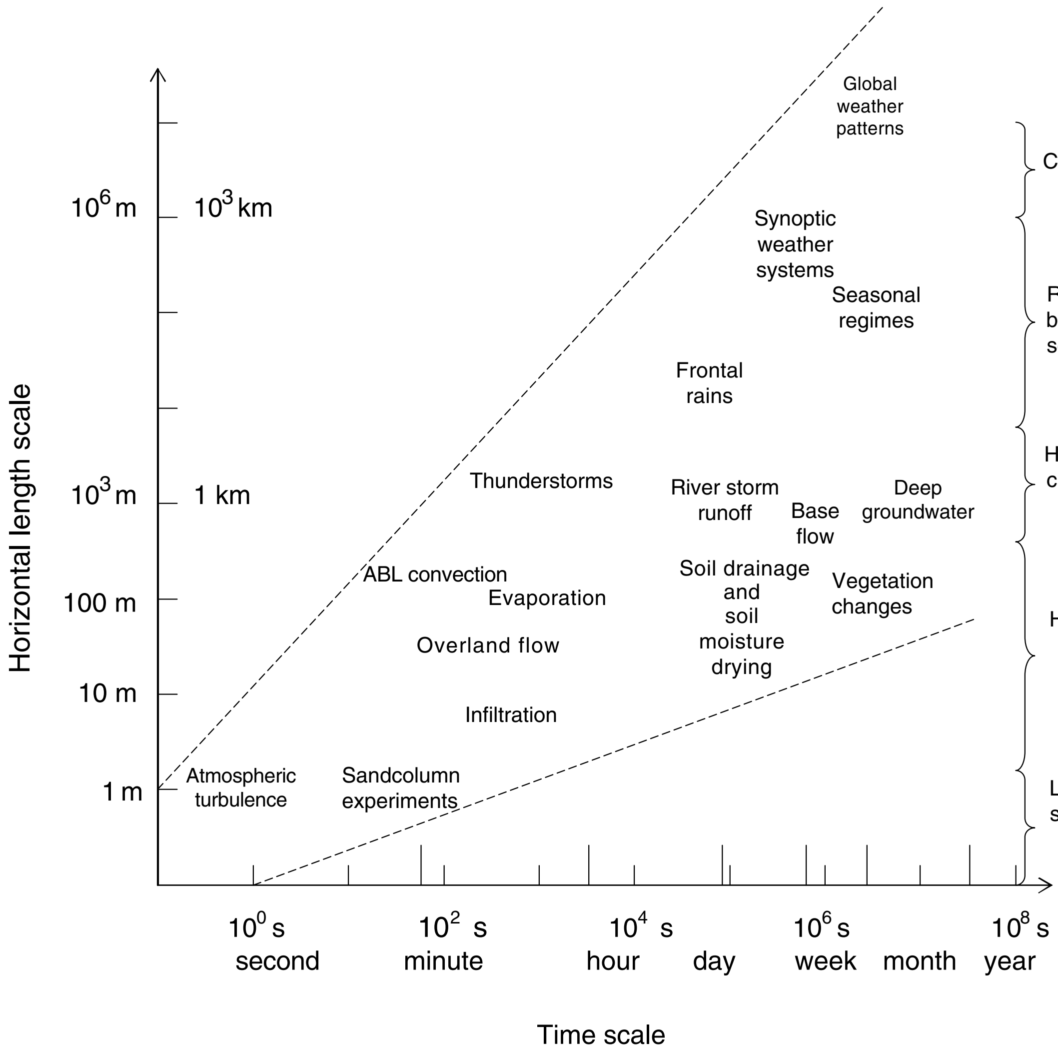

Figure 1.2 Approximate ranges of spatial and temporal scales of some common physical processes that are relevant in hydrology.

All this suggests that a sound criterion to distinguish, in principle at least, one approach from another may be the spatial scale at which the internal mechanisms are parameterized. For example, Newton's equation for viscous shear stress (see Equation (1.12) below) is a parameterization in terms of variables typically at the millimeter to centimeter scale; however, it reflects momentum exchanges at molecular scales, which are orders of magnitude smaller. The hydraulic conductivity is a parameter at the so-called Darcy scale (see Chapter 8), namely a scale somewhere intermediate between the Newtonian viscosity (or Navier–Stokes) scale for water and air inside the soil pores, on the one hand, and the field scales for infiltration and drainage, on the other. Several spatial scales are illustrated, with the corresponding characteristic temporal scales, in Figure 1.2 for some general types of water transport processes, as they have been considered in hydrologic studies.

Table 1.5 A common scale classification in the atmosphere (after Orlanski, 1975)

Nomenclature	Scale range
Micro γ	<20 m
Micro β	20–200 m
Micro α	200 m–2 km
Meso γ	2 km–20 km
Meso β	20 km–200 km
Meso α	200 km–2000 km
Macro γ	>2000 km

On the land surfaces of the Earth, the catchment or river basin sizes appear as scales of central importance. The terms *basin, catchment, watershed*, and *drainage area* are roughly synonymous and are often used interchangeably. A basin can be defined as all of the upstream area, which contributes to the open channel flow at a given point along a river. The size of the basin depends on the selection of the point in the river system under consideration. Usually this point is taken where the river flows into a large water body, such as a lake or the ocean, or where it changes its name as a tributary into a larger river. However, a basin or catchment can also be defined by any point along the river where the river flow is being measured. Basins are delineated naturally by the land-surface topography, and topographic ridges are usually taken as their boundaries; they can be considered as the natural conveyance systems for mass and energy on the land surfaces of the Earth. In meteorology, the concern is more on atmospheric motions and weather systems, and this has led to a somewhat different scale classification; an example of a commonly used classification is shown in Table 1.5.

To summarize, these observations indicate that, in deciding on a strategy to describe a hydrologic phenomenon, the relevant question is probably not so much whether a physical, a black box, or a conceptual approach should be used. Rather, it is more useful to determine what scales are appropriate for the available and measurable data, and for the problem at hand. In other words, what is the appropriate level of parameterization?

Spatial Variability and Effective Parameters

As mentioned above, a parameterization can be defined as a functional relationship between the variables describing the phenomenon in question. This relationship invariably contains one or more constant terms, reflecting material and fluid properties, and vegetational, geomorphic, geologic, and other physiographic features; these are called parameters and they are normally determined by experiment. Most hydrologic parameters tend to be highly variable in space. It stands to reason, therefore, that the experimental determination of any such space-dependent parameter must be carried out at the scale at which it is to be applied to describe the flow.

A second important issue is that any given parameterization is usually valid only over a certain finite range of spatial scales, and that the computational scale, that is the integration domain or the discretization of the equations, must lie within that range. Because the necessary data may be available only at a coarser resolution, in practical application, a parameterization may have to be applied at scales for which it was not intended originally and which are larger than permissible. This means that in such a case the spatial variability of the