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SCALE ISSUES IN HYDROLOGICAL MODELLING: A REVIEW

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ABSTRACT

A framework is provided for scaling and scale issues in hydrology. The first section gives some basic definitions. This is important as researchers do not seem to have agreed on the meaning of concepts such as scale or upscaling. 'Process scale', 'observation scale' and 'modelling (working) scale' require different definitions. The second section discusses heterogeneity and variability in catchments and touches on the implications of randomness and organization for scaling. The third section addresses the linkages across scales from a modelling point of view. It is argued that upscaling typically consists of two steps: distributing and aggregating. Conversely, downscaling involves disaggregation and singling out. Different approaches are discussed for linking state variables, parameters, inputs and conceptualizations across scales. This section also deals with distributed parameter models, which are one way of linking conceptualizations across scales. The fourth section addresses the linkages across scales from a more holistic perspective dealing with dimensional analysis and similarity concepts. The main difference to the modelling point of view is that dimensional analysis and similarity concepts deal with complex processes in a much simpler fashion. Examples of dimensional analysis, similarity analysis and functional normalization in catchment hydrology are given. This section also briefly discusses fractals, which are a popular tool for quantifying variability across scales. The fifth section focuses on one particular aspect of this holistic view, discussing stream network analysis. The paper concludes with identifying key issues and gives some directions for future research.

KEY WORDS Scale Scaling Aggregation Effective parameters Distributed modelling Dimensional analysis Similarity Fractals Stream network analysis Geomorphologic unit hydrograph

INTRODUCTION

This review is concerned with scale issues in hydrological modelling, with an emphasis on catchment hydrology.

Hydrological models may be either predictive (to obtain a specific answer to a specific problem) or investigative (to further our understanding of hydrological processes) (O'Connell, 1991; Grayson *et al.*, 1992). Typically, investigative models need more data, are more sophisticated in structure and estimates are less robust, but allow more insight into the system behaviour. The development of both types of model has traditionally followed a set pattern (Mackay and Riley, 1991; O'Connell, 1991) involving the following steps: (a) collecting and analysing data; (b) developing a conceptual model (in the researcher's mind) which describes the important hydrological characteristics of a catchment; (c) translating the conceptual model

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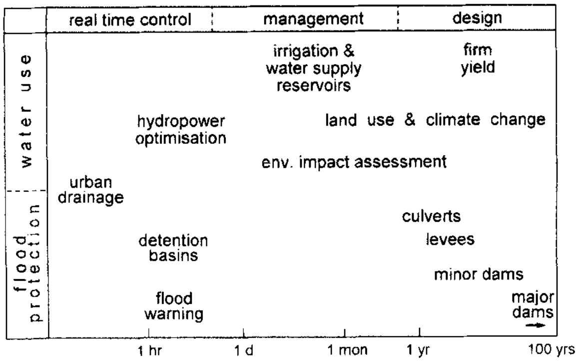

Figure 1. Problem solutions required at a range of time-scales

into a mathematical model; (d) calibrating the mathematical model to fit a part of the historical data by adjusting various coefficients; (e) and validating the model against the remaining historical data set.

If the validation is not satisfying, one or more of the previous steps needs to be repeated (Gutknecht, 1991a). If, however, the results are sufficiently close to the observations, the model is considered to be ready for use in a predictive mode. This is a safe strategy when the conditions for the predictions are similar to those of the calibration/validation data set (Bergström, 1991). Unfortunately, the conditions are often very different, which creates a range of problems. These are the thrust of this paper.

Conditions are often different in their space or time *scale*. The term *scale* refers to a characteristic time (or length) of a process, observation or model. Specifically, processes are often observed and modelled at short-time scales, but estimates are needed for very long time-scales (e.g. the life time of a dam). Figure 1 gives some examples of real-time control, management and design for which estimates of hydrological models are required. The associated time-scales range from minutes to hundreds of years. Similarly, models and theories developed in small space-scale laboratory experiments are expected to work at the large scale of catchments. Conversely, sometimes large-scale models and data are used for small-scale predictions. This invariably involves some sort of extrapolation, or equivalently, transfer of information across scales. This transfer of information is called *scaling* and the problems associated with it are *scale issues*.

In the past few years scale issues in hydrology have increased in importance. This is partly due to increased environmental awareness. However, there is still a myriad of unresolved questions and problems. Indeed, '... the issue of the linkage and integration of formulations at different scales has not been addressed adequately. Doing so remains one of the outstanding challenges in the field of surficial processes' (NRC, 1991: 143).

Scale issues are not unique to hydrology. They are important in a range of disciplines such as: meteorology and climatology (Haltiner and Williams, 1980; Avissar, 1995; Raupach and Finnigan, 1995); geomorphology (de Boer, 1992); oceanography (Stommel, 1963); coastal hydraulics (deVriend, 1991); soil science (Hillel and Elrick, 1990); biology (Haury *et al.*, 1977); and the social sciences (Dovers, 1995). Only a few papers have attempted a review of scale issues in hydrology. The most relevant papers are Dooge (1982; 1986), Klemeš (1983), Wood *et al.* (1990), Beven (1991) and Mackay and Riley (1991). Rodríguez-Iturbe and Gupta (1983) and Gupta *et al.* (1986a) provide a collection of papers related to scale issues and Dozier (1992) deals with aspects related to data.

This paper attempts to provide a framework for scaling and scale issues in hydrology. The first section gives some basic definitions. This is important as researchers do not seem to have agreed on what notions such as scale or upscaling exactly mean. The second section discusses heterogeneity and variability in catchments, which is indeed what makes scale issues so challenging. The third section addresses the linkages across scales from a modelling point of view. Specifically, different approaches are discussed for linking state variables, parameters, inputs and conceptualizations across scales. This section also deals with distributed parameter models, which are one way of linking conceptualizations across scales. The fourth section addresses the linkages across scales from a more holistic perspective, dealing with dimensional analysis and similarity concepts. The main difference to the modelling point of view is that dimensional analysis and similarity concepts deal with complex processes in a much simpler fashion. This section also briefly discusses fractals, which are a popular tool for quantifying variability across scales. The fifth section focuses on one particular aspect of this holistic view discussing stream network analysis. The paper concludes with identifying key issues and gives some directions for future research.

THE NOTION OF SCALES AND DEFINITIONS

Hydrological processes at a range of scales

Hydrological processes occur at a wide range of scales, from unsaturated flow in a 1 m soil profile to floods in river systems of a million square kilometres; from flashfloods of several minutes duration to flow in aquifers over hundreds of years. Hydrological processes span about eight orders of magnitude in space and time (Klemeš, 1983).

Figure 2 attempts a classification of hydrological processes according to typical length and time scales. Shaded regions show characteristic time–length combinations of hydrological activity (variability). This type of diagram was first introduced by Stommel (1963) for characterizing ocean dynamics and was later adopted by Fortak (1982) to atmospheric processes. Since then it has been widely used in the atmospheric sciences (e.g. Smagorinsky, 1974; Fortak, 1982). The shaded regions in Figure 2 can be thought of as regions of spectral power (in space and time) above a certain threshold. Stommel (1963: 572) noted, 'It is convenient to depict these different components of the spectral distribution of sea levels on a diagram (Stommel's Figure 1) in which the abscissa is the logarithm of period, P in seconds, and the ordinate is the logarithm of horizontal scale, L in centimeters. If we knew enough we could plot the spectral

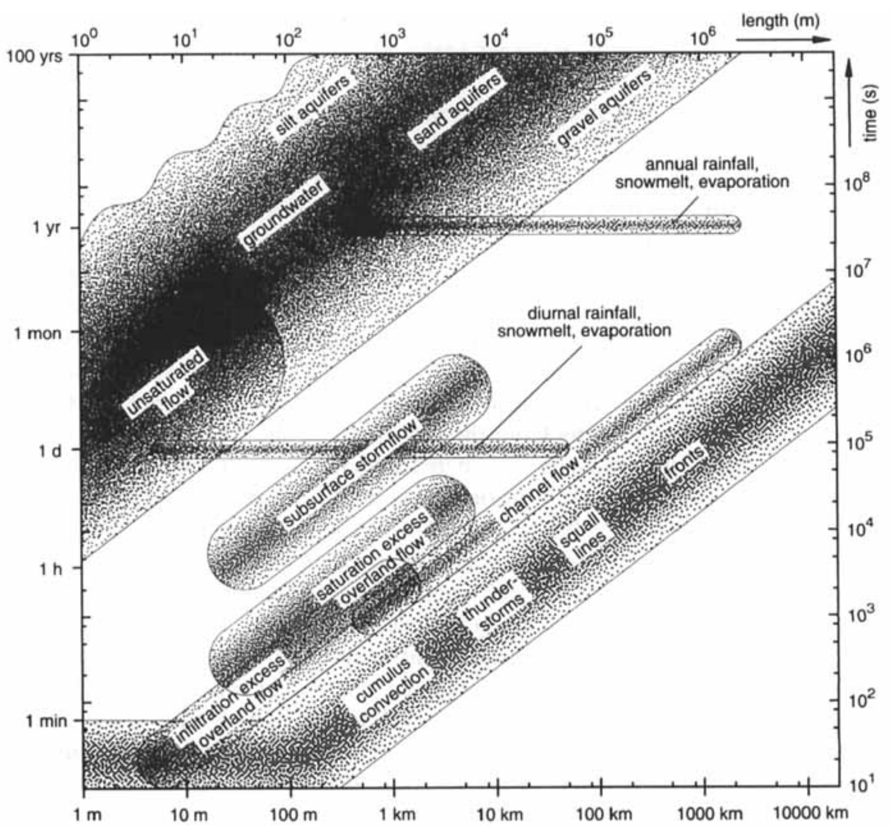

Figure 2. Hydrological processes at a range of characteristic space–time scales. Based on Orlanski (1975), Dunne (1978), Fortak (1982) and Anderson and Burt (1990) with additional information from the authors

distribution quantitatively by contours on this period-wavelength plane.' In a hydrological context, the shaded regions for a particular process in Figure 2 could be determined, at least conceptually, by plotting a characteristic time-scale (e.g. scale of maximum spectral power of a discharge record or, alternatively, response time of a catchment) versus a characteristic length scale (e.g. square root of catchment area).

Figure 2 is based on both data and heuristic considerations. Precipitation is one of the forcings driving the hydrological cycle. Precipitation phenomena range from cells (associated with cumulus convection) at scales of 1 km and several minutes, to synoptic areas (frontal systems) at scales of 1000 km and more than a day (Austin and Houze, 1972; Orlanski, 1975). Many hydrological processes operate — in response to precipitation — at similar length scales, but the time-scales are delayed. The time delay increases as the water passes through the subsurface and clearly depends on the dominant runoff mechanisms (Pearce *et al.*, 1986). Consider, for example, a small catchment of, say, 1 km^2 size. Infiltration excess (i.e. Horton overland flow) response (such as often occurs in an arid climate during high rainfall intensities) is very fast (<30 minutes). Saturation excess (i.e. saturation overland flow) response (e.g. in humid climates and thin soils) is typically slower because the building up of a saturated layer delays the runoff response. Subsurface stormflow is often significantly slower, with response times of a day or longer for the same catchment size. Finally, groundwater-controlled flows are associated with time-scales from months to hundreds of years (Dunne, 1978; 1983; Zuidema, 1985; Anderson and Burt, 1990).

As in the case of atmospheric processes, different hydrological processes occur at different length scales (Figure 2). Runoff generation associated with rainfall intensities exceeding infiltration capacities (which produces infiltration excess/Horton overland flow) is a 'point phenomenon' and can, as such, be defined at a very small length scale. Saturation excess runoff (i.e. saturation overland flow) is an integrating process and needs a certain minimum catchment area to be operative. This is because, typically, the main mechanism for raising the groundwater table (which in turn produces saturation overland flow) is *lateral* percolation above an impeding horizon. Also, subsurface stormflow needs a certain minimum catchment area to be operative. Channel flow typically occurs at larger scales above a channel initiation area up to the length scales of the largest river basins.

Figure 2 suggests a roughly constant ratio of characteristic length and time-scales for a given process over a range of scales. This ratio is termed the *characteristic velocity* (Haltiner and Williams, 1980; Blöschl *et al.*, 1995). For atmospheric processes this characteristic velocity is of the order of 10 m/s, for channel flow it is 1 m/s and for subsurface stormflow it is less than, say, 0·1 m/s. In Figure 2, the slopes have been selected as flatter than 1:1, which means a slight increase in the characteristic velocity with scale. This is consistent with Orlanski's (1975) definition of atmospheric scales, a slight tendency of channel flow velocities to increase with catchment scale (Leopold and Maddock, 1953) and the response times for catchments of different sizes (Anderson and Burt, 1990). It is speculated that, physically, this may be related to slightly decreased resistances (in a general sense) with scale. Also, the runoff response of a catchment is the combined effect of a number of processes. The combined characteristic velocity will reflect the relative contributions of individual processes. Specifically, when moving from the local scale to the regional scale, the increasing importance of channel flow will, typically, translate into increasing characteristic velocities with scale. In Figure 2 this means that the trace of the space–time relationship of combined processes (e.g. unsaturated flow, subsurface stormflow, channel flow) will be flatter than that of the individual processes shown.

Another forcing of the hydrological cycle is solar radiation. Consequently, a number of hydrological processes show a clear diurnal and annual cycle. Such processes include evaporation, snowmelt, and — depending on climate — precipitation (Gutknecht, 1993).

One of the striking features of Figure 2 is a certain relationship between typical length and time-scales for a given process. Small length scales tend to be associated with small time-scales and the same applies to large length and time-scales. Indeed, by looking at the plot of a hydrograph (with the time-scale given) we can roughly estimate the size of the catchment associated with it. For example, a slim and peaky hydrograph suggests a small catchment. Because of the relationship between length and time-scales of hydrological processes, we often just refer to 'the scale' of a process (either small or large) and implicitly assume that this relates to both length and time. However, when looking at processes in more detail, this is not always so. For example, unsaturated flow at the site scale (1 m) can be associated with time-scales

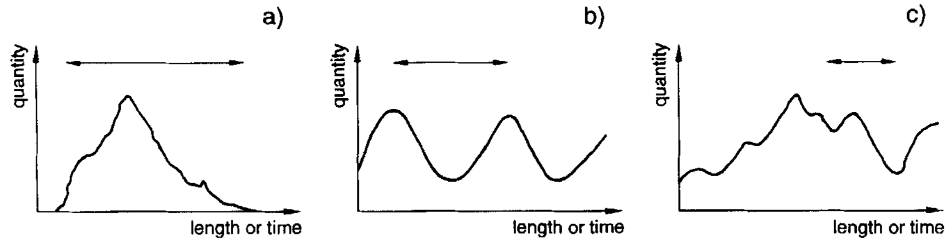

Figure 3. Three alternative definitions of process scale in space l (and time t). (a) Spatial extent (duration); (b) space (time) period; and (c) integral scale or correlation length (time)

from hours to weeks. To examine the scale of a process more closely, we need to first define, exactly, the notion of 'scale'. Specifically, we will first discuss the 'process scale' and then the 'observation scale'. The 'process scale' is the scale that natural phenomena exhibit and is beyond our control. On the other hand, we are free to choose the 'observation scale', within the constraints of measurement techniques and logistics.

Process scale

Characteristic *time-scales* of a hydrological process can be defined as (Figure 3): (a) the lifetime (= duration) (for intermittent processes such as a flood); (b) the period (cycle) (for a periodic process such as snowmelt); and (c) the correlation length (= integral scale) (for a stochastic process exhibiting some sort of correlation) (Haltiner and Williams, 1980; Dooge, 1982; 1986; Klemeš, 1983; Dagan, 1986; Stull, 1988). Similarly, characteristic *space scales* can be defined either as spatial extent, period or integral scale, depending on the nature of the process. Specifically, for a random function $w(\boldsymbol{x})$ that is stationary such that its covariance $C_w(\boldsymbol{x}_1, \boldsymbol{x}_2)$ depends on $\boldsymbol{r} = \boldsymbol{x}_1 - \boldsymbol{x}_2$ rather than $\boldsymbol{x}_1$ and $\boldsymbol{x}_2$, the integral scale is defined as

$$I_w = \int \frac{C_w(\boldsymbol{r})\mathrm{d}\boldsymbol{r}}{\sigma_w^2} \tag{1}$$

where $\boldsymbol{x}$ is either space or time, $\boldsymbol{r}$ is space (or time) lag and σ_w^2 is the variance. In other words, the integral scale refers to the average distance (or time) over which a property is correlated (Dagan, 1986). 'Correlation length' has two main usages: the first is identical with that of the integral scale, i.e. the average distance of correlation. The second refers to the maximum distance of correlation (de Marsily, 1986). Clark (1985) discusses the relationship between a number of different scale definitions.

The various definitions for scale (lifetime, period, correlation length) are often used interchangeably and it is not always clear, in a particular case (of, say, a stochastic *and* periodic phenomenon), which one is used. One justification for doing this is the relatively small difference between the above definition compared with the total range in scale of eight orders of magnitude. Indeed, the term 'scale' refers to a rough indication of the order of magnitude rather than to an accurate figure. In flood frequency analysis recurrence probabilities are sometimes thought of as a kind of scale. For example, a 100 year flood refers to a recurrence probability of 0·01 in any one year. This definition of scale is related to the interval *between* events (see crossing theory and the definition of runs, e.g. Yevjevich, 1972: 174).

Some hydrological processes exhibit (one or more) preferred scales, i.e. certain length (or time) scales are more likely to occur than others. These preferred scales are also called the *natural scales*. In a power spectrum representation preferred scales appear as peaks of spectral variance (Padmanabhan and Rao, 1988). Scales that are less likely to occur than others are often related to as a *spectral gap*. This name derives from the power spectrum representation in which the spectral gap appears as a minimum in spectral variance (Stull, 1988; 33). The existence of a spectral gap is tantamount to the existence of a *separation of scales* (Gelhar, 1986). Separation of scales refers to a process consisting of a small-scale (or fast) component superimposed on a much larger (or slow) component with a minimum in spectral variance in between. In meteorology, for example, microscale turbulence and synoptic scale processes are often separated by a spectral gap which manifests itself in a minimum in the power spectrum of windspeeds at the frequency of 1/hour.

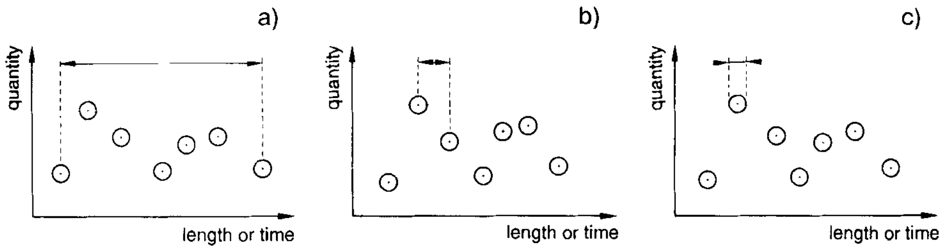

Figure 4. Three alternative definitions of measurement scale in space l (and time t). (a) Spatial (temporal) extent; (b) spacing (= resolution); and (c) integration volume (time constant)

In the time domain, many hydrological processes (such as snowmelt) exhibit preferred time-scales of one day and one year with a spectral gap in between. Clearly, this relates to the periodicity of solar radiation (Figure 2). In the space domain, there is no clear evidence for the existence of preferred scales. Precipitation, generally, does not seem to exhibit preferred scales and/or spectral gaps (e.g. Gupta and Waymire, 1993). Wood *et al.* (1988) suggested that catchment runoff may show a spectral gap at 1 km^2 catchment area. More recent work by Blöschl *et al.* (1995), however, indicates that both the existence and size of a spectral gap is highly dependent on specific catchment properties and climatic conditions. Also, runoff data (Woods *et al.*, 1995) do not support the existence of a spectral gap in runoff.

Observation scale

The definition of the 'observation scale' is related to the necessity of a finite number of samples. Consequently, 'observation scale' in space and time can be defined (Figure 4) as: (a) the spatial (temporal) extent (= coverage) of a data set; (b) the spacing (= resolution) between samples; or (c) the integration volume (time) of a sample.

The integration volume may range from 1 dm^3 for a soil sample to many square kilometres (i.e. the catchment size) for discharge measurements. The integration time (i.e. the time constant) is often a property of a particular instrument. Typically, space resolutions are much poorer than time resolutions in hydrology. This is seen as one of the major obstacles to fast progress in scale-related issues. Obviously, the particular observation scale chosen dictates the type of the instrumentation, from detailed sampling of a soil profile to global coverage by satellite imagery (Dozier, 1992).

Process versus observation scale

Ideally, processes should be observed at the scale they occur. However, this is not always feasible. Often the interest lies in large-scale processes while only (small-scale) point samples are available. Also, hydrological processes are often simultaneously operative at a range of scales.

Cushman (1984; 1987), in a series of papers, discussed the relationship between process and observation scale and pointed out that sampling involves filtering. Figure 5 gives a more intuitive picture of the effect of

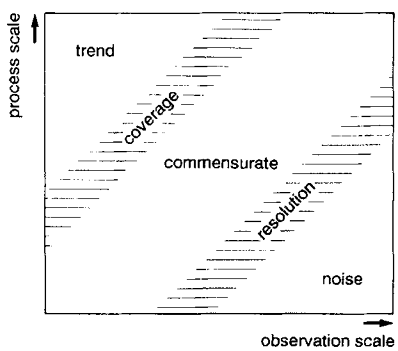

Figure 5. Process scale versus observation scale

sampling: processes larger than the coverage appear as trends in the data, whereas processes smaller than the resolution appear as noise. Specifically, the highest frequency that can be detected by a given data set of spacing d is given by the Nyquist frequency f_n (Jenkins and Watts, 1968; Russo and Jury, 1987)

$$f_n = \frac{1}{2d} \tag{2}$$

Modelling (working) scale

Yet another scale is the modelling (or working) scale. The modelling scales generally agreed upon within the scientific community are partly related to processes (Figure 2) and partly to the applications of hydrological models (Figure 1).

In space, typical modelling scales are (Dooge, 1982; 1986):

the local scale (1 m);
the hillslope (reach) scale (100 m);
the catchment scale (10 km);
and the regional scale (1000 km).

In time, typical modelling scales are:

the event scale (1 day);
the seasonal scale (1 yr);
and the long-term scale (100 yrs).

Unfortunately, more often than not, the modelling scale is much larger or much smaller than the observation scale. To bridge that gap, 'scaling' is needed.

Definition of upscaling, downscaling and regionalization

To scale, literally means 'to zoom' or to reduce/increase in size. In a hydrological context, *upscaling* refers to transferring information from a given scale to a larger scale, whereas *downscaling* refers to transferring information to a smaller scale (Gupta *et al.*, 1986a). For example, measuring hydraulic conductivity in a borehole and assuming it applies to the surrounding area involves upscaling. Also, estimating a 100 year flood from a 10 year record involves upscaling. Conversely, using runoff coefficients derived from a large catchment for culvert design on a small catchment involves downscaling (Mein, 1993). *Regionalization*, on the other hand, involves the transfer of information from one catchment (location) to another (Kleeberg, 1992). This may be satisfactory if the catchments are similar (in some sense), but error-prone if they are not (Pilgrim, 1983). One of the factors that make scaling so difficult is the heterogeneity of catchments and the variability of hydrological processes.

NATURE OF HETEROGENEITY AND VARIABILITY IN SPACE AND TIME

Heterogeneity at a range of scales

Natural catchments exhibit a stunning degree of heterogeneity and variability in both space and time. For example, soil properties and surface conditions vary in space, and vegetation cover, moisture status and flows also vary in time. The term 'heterogeneity' is typically used for media properties (such as hydraulic conductivity) that vary in space. The term 'variability' is typically used for fluxes (e.g. runoff) or state variables (e.g. soil moisture) that vary in space and/or time. Heterogeneity and variability manifest themselves at a range of scales.

Figure 6a illustrates subsurface spatial heterogeneity in a catchment. At the local scale, soils often exhibit macropores such as cracks, root holes or wormholes. These can transport the bulk of the flow with a minimum contribution of the soil matrix (Beven, 1981; Germann, 1990). For the macropore flow to become operative, certain thresholds in precipitation intensity and anticedent moisture may need to be met (Kneale and White, 1984; Germann, 1986). At the hillslope scale, preferential flow may occur through

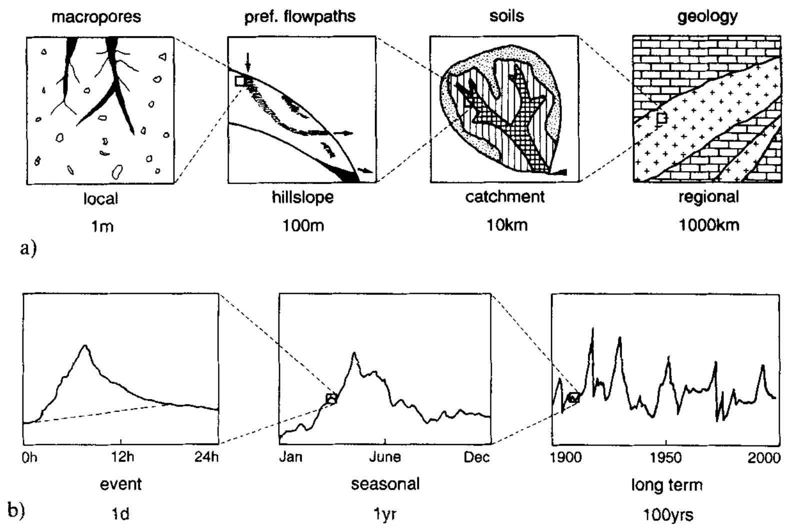

Figure 6. Heterogeneity (variability) of catchments and hydrological processes at a range of (a) space scales and (b) time-scales

high conductivity layers and pipes (Jones, 1987; Chappell and Ternan, 1992) and water may exfiltrate to the surface as return flow (Dunne, 1978). Figure 7 shows an example of flow along such a high conductivity layer in a catchment in the Austrian Alps. The location of the springs in Figure 7 indicates that the layer is parallel to the surface. Heterogeneity at the catchment scale (Figure 6a) may relate to different soil types and properties. Typically, valley floors show different soils as hillslopes and ridges (Jenny, 1980). At the regional scale, geology is often dominant through soil formation (parent material) and controls on the stream network density (von Bandat, 1962; Strahler, 1964).

Similarly, variability in time is present at a range of scales (Figure 6b). At the event scale the shape of the runoff wave is controlled by the characteristics of the storm and the catchment. At the seasonal scale, runoff is dominated by physioclimatic characteristics such as the annual cycle in precipitation, snowmelt and evaporation. Finally, in the long term, runoff may show the effects of long-term variability of precipitation (Mandelbrot and Wallis, 1968), climate variability (Schönwiese, 1979), geomorphological processes (Anderson, 1988) and anthropogenic effects (Gutknecht, 1993).

Types of heterogeneity and variability

The classification of heterogeneity (variability) is useful as it has a major bearing on predictability and scaling relationships (Morel-Seytoux, 1988). In fact, such a classification may serve as a predictive framework (White, 1988). Here, we will follow the suggestion of Gutknecht (1993), which is consistent with the definitions of 'scale' given earlier (Figure 3).

Hydrological processes may exhibit one or more of the following aspects: (a) *discontinuity* with discrete zones (e.g. intermittency of rainfall events; geological zones) — within the zones, the properties are relatively uniform and predictable, whereas there is disparity between the zones; (b) *periodicity* (e.g. the diurnal or annual cycle of runoff), which is predictable; and (c) *randomness*, which is not predictable in detail, but predictable in terms of statistical properties such as the probability density function.

At this stage it might be useful to define a number of terms related to heterogeneity and variability. *Disorder* involves erratic variation in space or time similar to randomness, but it has no probability aspect. Conversely, *order* relates to regularity or certain constraints (Allen and Starr, 1982). *Structure*, on the other hand, has two meanings. The first, in a general sense, is identical with order (e.g. 'storm structure'; Austin and Houze, 1972; geological 'structural facies', Anderson, 1989). The second, more specific, meaning is used in stochastic approaches and refers to the moments of a distribution such as mean, variance and correlation length (Hoeksema and Kitanidis, 1985). 'Structural analysis', for

Figure 7. Preferential flow along a high conductivity layer. Löhnersbach catchment in the Austrian Alps. Photo courtesy of R. Kirnbauer and P. Haas

example, attempts to estimate these moments (de Marsily, 1986). *Organization* is often used in a similar way to 'order', but tends to relate to a more complex form of regularity. Also, it is often closely related to the function and the formation (genesis) of the system (Denbigh, 1975). These aspects are illustrated in more detail in the ensuing discussion.

Organization

Catchments and hydrological processes show organization in many ways, which include the following examples. (a) Drainage networks embody a 'deep sense of symmetry' (Rodríguez-Iturbe, 1986). This probably reflects some kind of general principle such as minimum energy expenditure (Rodríguez-Iturbe *et al.*, 1992b) and adaptive landforms (Willgoose *et al.*, 1991). Also, Rodríguez-Iturbe (1986) suggested that Horton's laws (Horton, 1945; Strahler, 1957) are a reflection of this organization. (b) Geological facies (Walker, 1984) are discrete units of different natures which are associated with a specific formational process (Anderson, 1989; 1991; Miall, 1985). These units are often 'organized' in certain sequences. For example, in a glacial outwash sequence, deposits close to the former ice fronts mostly consist of gravel, whereas those further away may consist mainly of silt and clay. (c) Austin and Houze (1972) showed that precipitation patterns are organized in clearly definable units (cells, small mesoscale areas, large mesoscale areas, synoptic areas), which may be recognized as they develop, move and dissipate. (d) Soils tend to develop in response to state factors (i.e. controls) such as topography (Jenny, 1941; 1980). Different units in a catchment (e.g. 'nose', 'slope' and 'hollow'; Hack and Goodlett, 1960; England and Holtan, 1969; Krasovskaia, 1982) may have a different function and are typically formed by different processes. Soil catenas (i.e. soil sequences along a slope) are a common form of organization of soils in catchments (Milne, 1935; Moore *et al.*, 1993a; Gessler *et al.*, 1993).

Organization versus randomness

Randomness is essentially the opposite of organization (Dooge, 1986). Given that 'randomness is the property that makes statistical calculations come out right' (Weinberg, 1975: 17) and the high degree of organization of catchments illustrated in the preceding sections, care must be exercised when using stochastic methods. It is interesting to follow a debate between field and stochastic hydrogeology researchers on exactly this issue: Dagan (1986) put forward a stochastic theory of groundwater flow and transport and Neuman (1990) proposed a universal scaling law for hydraulic conductivity, both studies drawing heavily on the randomness assumption. Two comments prepared by Williams (1988) and Anderson (1991), respectively, criticized this assumption and emphasized the presence of organized discrete units as formed by geological processes. Specifically, Williams (1988) pointed out that the apparent disorder is largely a consequence of studying rocks through point measurements such as boreholes. This statement is certainly also valid in catchment hydrology. It is clear that for 'enlightened scaling' (Morel-Seytoux, 1988), the identification of organization is a key factor. The importance of organization at the hillslope and catchment scale has been exemplified by Blöschl *et al.* (1993), who simulated runoff hydrographs based on organized and random spatial patterns of hydrological quantities. Although the covariance structure of the organized and the random examples were identical, the runoff responses were vastly different. Similar effects have been shown by Kupfersberger and Blöschl (1995) for groundwater flow. It is also interesting to note that hydrological processes often show aspects of both organization and randomness (Gutknecht, 1993; Blöschl *et al.*, submitted).

LINKAGES ACROSS SCALES FROM A MODELLING PERSPECTIVE

Framework

Earlier in this paper, the term 'scaling' was defined as transferring information across scales. What, precisely, does information mean in this context? Let $g\{s;\theta;i\}$ be some small-scale conceptualization of hydrological response as a function of state variables s, parameters θ and inputs i, and let $G\{S;\Theta;I\}$ be the corresponding large-scale description. Now, the information scaled consists of state variables, para-

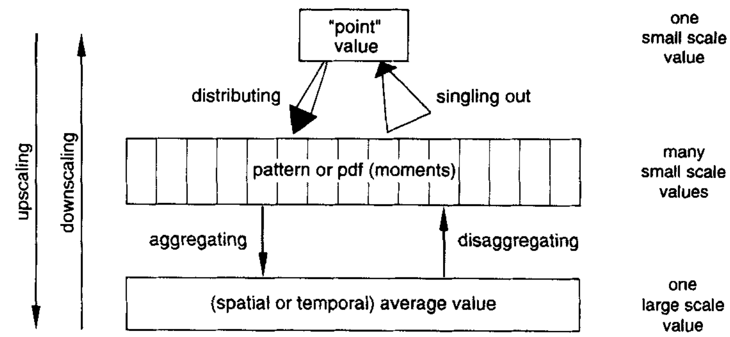

Figure 8. Linkages across scales as a two-step procedure

meters and inputs as well as the conceptualization itself

$$s \leftrightarrow S; \quad \theta \leftrightarrow \Theta; \quad i \leftrightarrow I; \quad g\{s;\theta;i\} \leftrightarrow G\{S;\Theta;I\} \tag{3}$$

However, in practice, often only one piece of information out of this list is scaled and the others are, implicitly, assumed to hold true at either scale. It is clear that this is not always appropriate. Also, the definition of a large-scale parameter Θ is not always meaningful, even if θ and Θ share the same name. The same can be the case for state variables. For example, it may be difficult (if not impossible) to define a large-scale hydraulic gradient in an unsaturated structured soil.

Upscaling, typically, consists of two steps. For illustration, consider the problem of estimating catchment rainfall from one rain gauge (or a small number of rain gauges), i.e. upscaling rainfall from a dm^2 scale to a km^2 scale. The first step involves *distributing* the small-scale precipitation over the catchment (e.g. as a function of topography). The second step consists of *aggregating* the spatial distribution of rainfall into one single value (Figure 8). Conversely, downscaling involves *disaggregating* and *singling out*. In some instances, the two steps are collapsed into one single step, as would be the case of using, say, station weights in the example of upscaling precipitation. However, in other instances only one of the steps is of interest. This has led to a very unspecific usage of the term 'upscaling' in published work referring to either distributing, aggregating, or both. In this paper we will use the terms as defined in Figure 8.

Scaling can be performed either in a deterministic or a stochastic framework. In a deterministic framework, distributing small-scale values results in a spatial (or temporal) pattern which is aggregated into one 'average' value. In a stochastic framework, distributing results in a distribution function (and covariance function), often characterized by its moments, which is subsequently aggregated. The stochastic approach has a particular appeal as the detailed pattern is rarely known and distribution functions can be derived

Table I. Examples for linkages across scales. Abbreviations: swe = snow water equivalent

	State variables	Parameters	Inputs	Conceptualizations
Distributing	Wetness index swe $= f$(terrain)	Kriging conductivities Soil type mapping	Spline interpolation of rainfall	Distributed model
Singling out	Trivial	Trivial	Trivial	Trivial
Aggregation	Mostly trivial	Effective conductivity = geometric mean	Trivial	Perturbation approach
Disaggregation	Wetness index and satellite data		Mass curves Daily $\rightarrow$ hourly rainfall	
Upscaling in one step	Snow index stations		Thiessen method depth–area curves	
Downscaling in one step		Runoff coefficient for culvert design		

more readily. On the other hand, deterministic methods have more potential to capture the organized nature of catchments, as discussed earlier in this paper.

Methods of scaling greatly depend on whether state variables, parameters, inputs or conceptualizations are scaled. Although parameters often need to be scaled in the context of a particular model or theory, inputs and state variables can in many instances be treated independently of the model. Table I gives some examples that are of importance in hydrology. Distributing information is usually addressed by some sort of interpolation scheme such as kriging. Singling out is always trivial as it simply involves selecting a subset of a detailed spatial (or temporal) pattern that is known. Aggregating information is trivial or mostly trivial for state variables (e.g. soil moisture) and inputs (e.g. rainfall) as the aggregated value results from laws such as the conservation of mass. It is not trivial, however, to aggregate model parameters (e.g. hydraulic conductivity) as the aggregated value depends on the interplay between the model and the model parameters. Disaggregating information is often based on indices such as the wetness index. Methods for upscaling/downscaling in one step are often based on empirical regression relationships. Finally, scaling conceptualizations refers to deriving a model structure from a model or theory at another scale. Recall that no scaling is involved for a model that is formulated directly at the scale at which inputs are available and outputs are required. Some of the examples in Table I are discussed in more detail in the following sections.

Distributing information

Distributing information over space or time invariantly involves some sort of interpolation. Hydrological measurements are, typically, much coarser spaced in space than in time, so most interpolation schemes refer to the space domain.

The classical interpolation problem in hydrology is the spatial estimation of rainfall from rain gauge measurements. A wide variety of methods has been designed, including: the isohyetal method; optimum interpolation/kriging (Matheron, 1973; Journel and Huijbregts, 1978; Deutsch and Journel, 1992); spline interpolation (Creutin and Obled, 1982; Hutchinson, 1991); moving polynomials; inverse distance; and others. The various methods have been compared in Creutin and Obled (1982), Lebel *et al.* (1987) and Tabios and Salas (1985), and a review is given in Dingman (1994: 120).

Other variables required for hydrological modelling include hydraulic conductivity (de Marsily, 1986), climate data (Hutchinson, 1991; Kirnbauer *et al.*, 1994) and elevation data (Moore *et al.*, 1991), for which similar interpolation methods are used.

In many instances the supports (i.e. measurements) on which the interpolation is based are too widely spaced and the natural variability in the quantity of interest is too large for reliable estimation. One way to address this problem is to correlate the quantity of interest to an auxiliary variable (i.e. a *covariate* or *surrogate*), whose spatial distribution can more readily be measured. The spatial distribution of the quantity is then inferred from the spatial distribution of the covariate.

In catchment hydrology topography is widely used as a covariate because it is about the only information known in a spatially distributed fashion. Precipitation has a tendency to increase with elevation on an event (Fitzharris, 1975) and seasonal (Lang, 1985) scale, but this is not necessarily so for hourly and shorter scales (Obled, 1990). A number of interpolation schemes have been suggested that explicitly use elevation, ranging from purely statistical (Jensen, 1989) to dynamic (Leavesley and Hay, 1993) approaches. Similarly, snow characteristics, though highly variable at a range of space scales, are often related to terrain (Golding, 1974; Woo *et al.*, 1983a; 1983b). For example, Blöschl *et al.* (1991) suggested an interpolation procedure based on elevation, slope and curvature. The coefficients relating snow water equivalent to elevation were based on a best fit to snow course data, whereas the other coefficients were derived from qualitative considerations.

Topographic information has also been used to estimate the spatial distribution of soil moisture. The topographic wetness index was first developed by Beven and Kirkby (1979) and O'Loughlin (1981) to predict zones of saturation. The wetness index of Beven and Kirkby (1979) is based on four assumptions:

1. The lateral subsurface flow-rate is assumed to be proportional to the local slope $\tan\beta$ of the terrain. This implies kinematic flow, small slopes ($\tan\beta \approx \sin\beta$) and that the water-table is parallel to the topography.

2. Hydraulic conductivity is assumed to decrease exponentially with depth and storage deficit is assumed to be distributed linearly with depth (Beven, 1986).
3. Recharge is assumed to be spatially uniform.
4. Steady-state conditions are assumed to apply, so the lateral subsurface flow-rate is proportional to the recharge and the area a drained per unit contour length at point i.

Introducing the wetness index w_i at point i

$$w_i = \ln\left(\frac{\bar{T}\cdot a}{T_i \cdot \tan\beta}\right) \tag{4}$$

where T_i is the local transmissivity and $\bar{T}$ is its average in the basin, gives a simple expression for the storage deficit S_i at point i

$$S_i = \bar{S} + m\cdot(\bar{w} - w_i) \tag{5}$$

$\bar{S}$ and $\bar{w}$ are the averages of the storage deficit and the wetness index over the catchment, respectively, and m is an integral length which can be interpreted as the equivalent soil depth. For a detailed derivation, see Beven and Kirkby (1979), Beven (1986) and Wood *et al.* (1988). A similar wetness index has been proposed by O'Loughlin (1981;1986) which, however, makes no assumption about the shape of hydraulic conductivity with depth.

The wetness index w_i increases with increasing specific catchment area and decreasing slope gradient. Hence the value of the index is high in valleys (high specific catchment area and low slope), where water concentrates, and is low on steep hillslopes (high slope), where water is free to drain. For distributing the saturation deficit S_i over a catchment, Equation (5) can be fitted to a number of point samples. However, the predicitive power of the wetness index has not yet been fully assessed. Most comparisons with field data were performed in very small catchments. Moore *et al.* (1988) found that the wetness index [Equation (4)] explained 26–33% of the spatial variation in soil water content in a 7·5 ha catchment in NSW, Australia, whereas a study in a 38 ha catchment in Kansas (Ladson and Moore, 1992) suggested an explained variation of less than 10%. Burt and Butcher (1985) found that the wetness index explained 17–31% of the variation in the depth to saturation on a 1·4 ha hillslope in south Devon, UK. There also seems to be a dependence of the wetness index on the grid size used for its calculation, which may become important when using the wetness index in larger catchments (Moore *et al.*, 1993b; Vertessy and Band, 1993; Band and Moore, 1995).

A range of wetness indices has been suggested which attempts to overcome some of the limitations of Equation (4). For example, Barling *et al.* (1994) developed a quasi dynamic wetness index that accounts for variable drainage times since a prior rainfall event. This relaxes the steady-state assumption. Other developments include the effect of evaporation as controlled by solar radiation (e.g. Moore *et al.*, 1993c). A comprehensive review is given in Moore *et al.* (1991). Based on the rationale that in many landscapes pedogenesis of the soil catena (Milne, 1935) occurs in response to the way water moves through the landscape (Jenny, 1941; 1980; Hoosbeek and Bryant, 1992), terrain indices have also been used to predict soil attributes (McKenzie and Austin, 1993). For example, Moore *et al.* (1993a) predicted quantities such as A horizon thickness and sand content on a 5·4 ha toposequence in Colorado where the explained variances were about 30%. Gessler *et al.* (1993) reported a similar study in south-east Australia.

There is a range of other covariates in use for distributing hydrological information across catchments. Soil type or particle size distributions are the traditional indices to distribute soil hydraulic properties (Rawls *et al.*, 1983; Benecke, 1992; Williams *et al.*, 1992). General purpose soil maps are widely available in considerable detail, such as the STATSGO database developed by the US Department of Agriculture (Reybold and TeSelle, 1989). Unfortunately, the variation of soil parameters (such as hydraulic conductivity) within a particular soil class is often much larger than variations between different soils (Rawls *et al.*, 1983; McKenzie and MacLeod, 1989). Part of the problem is that the spatial scale of variation of hydraulic properties tends to be much smaller (<10 m) than the resolution of most soil maps (Bridges, 1982).

Alternatively, geophysical data have been used for estimating quantities of interest to hydrological modelling (e.g. Rubin *et al.*, 1992; Copty *et al.*, 1993). Mazac *et al.* (1985) discuss the factors affecting the relationship between the electrical and hydraulic properties of aquifers. For example, an increased clay content tends to decrease both the electrical resistivity and hydraulic conductivity. Kupfersberger and Blöschl (1995) showed that auxiliary (geophysical) data are particularly useful for identifying high hydraulic conductivity zones such as buried stream channels. There is also a range of indices supporting the interpolation of information related to vegetation. For example, Hatton and Wu (1995) used the leaf area index (LAI) for interpolating measurements of tree water use. Further covariates include satellite data, particularly for the spatial estimation of precipitation, snow properties, evapotranspiration and soil moisture (Engman and Gurney, 1991). However, a full review of this is beyond the scope of this paper.

An exciting new area is that of using indicators (Journel, 1986) to spatially distribute information. Indicators are binary variables which are often more consistent with the presence of zones of uniform properties (e.g. clay lenses or geological zones at a larger scale) than the ubiquitous log-normal distributions (Hoeksema and Kitanidis, 1985). They are also more consistent with the type and amount of information usually available. Indicator-based methods have been used to estimate conductivity in aquifers (Brannan and Haselow, 1993; Kupfersberger, 1994) and they have also been used to interpolate rainfall fields (Barancourt *et al.*, 1992). In the rainfall example, the binary patterns represent areas of rainfall/no rainfall.

Aggregating model parameters

The study of the aggregation of model parameters involves two questions. (a) Can the microscale equations be used to describe processes at the macroscale? (b) If so, what is the aggregation rule to obtain the macroscale model parameters with the detailed pattern or distribution of the microscale parameters given.

The macroscale parameters for use in the microscale equations are termed *effective parameters*. Specifically, an effective parameter refers to a single parameter value assigned to all points within a model domain (or part of the domain) such that the model based on the uniform parameter field will yield the same output as the model based on the heterogeneous parameter field (Mackay and Riley, 1991). Methods to address the questions (a) and (b) make use of this definition by matching the outputs of the uniform and heterogeneous systems. If an adequate match can be obtained, an effective parameter exists. The aggregation rule is then derived by relating the effective parameter to the underlying heterogeneous distribution. Methods used include analytical approaches (e.g. Gutjahr *et al.*, 1978), Monte Carlo simulations (e.g. Binley *et al.*, 1989) and measurements (e.g. Wu *et al.*, 1982). Effective parameters are of clear practical importance in distributed modelling, but aggregation rules are of more conceptual rather than of practical importance, either in a deterministic or a stochastic framework. In a deterministic framework, the aggregation rule yields an effective parameter over a certain domain with the pattern of the detailed parameter given. However, in practical applications there is rarely enough information on the detailed pattern available to use the aggregation rule in a useful way. In a stochastic framework, the aggregation rule yields the spatial moments of the effective parameter with the spatial moments of the microscale parameter given. This has been used to determine element spacings or target accuracy for calibration (Gelhar, 1986), but in practice other considerations such as data availability are often more important. Another limitation is the assumption of disordered media properties on which such approaches are often based (e.g. Rubin and Gómez-Hernández, 1990). If some sort of organization is present (e.g. preferential flow; Silliman and Wright, 1988), the same aggregation rules cannot be expected to hold.

What follows is a review of effective parameters and aggregation rules for a number of processes that are of importance in catchment hydrology. These include saturated flow, unsaturated flow, infiltration and overland flow.

The concept of effective hydraulic conductivity has been widely studied for saturated flow. Consider, in a deterministic approach, uniform (parallel flow lines) two-dimensional steady saturated flow through a block of porous medium made up of smaller blocks of different conductivities. It is easy to show that for an arrangement of blocks in series the effective conductivity equals the harmonic mean of the block values. Similarly, for an arrangement of blocks in parallel, the effective conductivity equals the arithmetic mean.

In a stochastic approach, the following results were found for steady saturated flow in an infinite domain without sinks (i.e. uniform flow).

(a) Whatever the spatial correlation and distribution function of conductivity and whatever the number of dimensions of the space, the average conductivity always ranges between the harmonic mean and the arithmetic mean of the local conductivities (Matheron, 1967, cited in de Marsily, 1986). Specifically, for the one-dimensional case, the effective conductivity K_{eff} equals the harmonic mean K_{H} as it is equivalent to an arrangement of blocks in series

$$\text{One-dimensional:} \qquad K_{\mathrm{eff}} = K_{\mathrm{H}} \tag{6}$$

(b) If the probability density function of the conductivity is log-normal (and for any isotropic spatial correlation), Matheron (1967) and Gelhar (1986) showed that for the two-dimensional case the effective conductivity K_{eff} equals the geometric mean K_{G}.

$$\text{Two-dimensional:} \qquad K_{\mathrm{eff}} = K_{\mathrm{G}} \tag{7}$$

Matheron also made the conjecture that for the three-dimensional case the effective conductivity K_{eff} is

$$\text{Three-dimensional:} \qquad K_{\mathrm{eff}} = K_{\mathrm{G}} \cdot \exp(\sigma^2_{\ln K}/6) \tag{8}$$

where $\sigma^2_{\ln K}$ is the variance of $\ln K$. This result is strongly supported by the findings of Dykaar and Kitanidis (1992) using a numerical spectral approach. Dykaar and Kitanidis (1992) also suggested that Equation (8) is more accurate than the small perturbation method (Gutjahr *et al.*, 1978) and the imbedding matrix method (Dagan, 1979) when the variances $\sigma^2_{\ln K}$ are large.

(c) If the probability distribution function is not log-normal (e.g. bimodal), both the numerical spectral approach and the imbedding matrix method seem to provide good results, whereas the small perturbation method is not applicable (Dykaar and Kitanidis, 1992).

Unfortunately, steady-state conditions and uniform flow are not always good assumptions (Rubin and Gómez-Hernández, 1990). For transient conditions, generally, no effective conductivities independent of time may be defined (Freeze, 1975; El-Kadi and Brutsaert, 1985). Similarly, for bounded domains and flow systems involving well discharges the effective conductivity is dependent on pumping rates and boundary conditions (Gómez-Hernández and Gorelick, 1989; Ababou and Wood, 1990; Neuman and Orr, 1993). It is interesting to note that the effective conductivity tends to increase with increasing dimension of the system for a given distribution of conductivity. This is consistent with intuitive reasoning: the higher the dimension the more degrees of freedom are available from which flow can 'choose' the path of lowest resistance. Hence, for a higher dimension, flow is more likely to encounter a low resistance path which tends to increase the effective conductivity.

For unsaturated flow in porous media, generally, there exists no effective conductivity that is a property of the medium only (Russo, 1992). Yeh *et al.* (1985) suggested that the geometric mean is a suitable effective parameter for certain conditions, but Mantoglou and Gelhar (1987) showed that the effective conductivity is heavily dependent on a number of variables such as the capillary tension head. In their examples, an increase in the capillary tension head from 0 to 175 cm translated into a decrease in the effective conductivity of up to 10 orders of magnitude. Mantoglou and Gelhar (1987) also demonstrated significant hysteresis in the effective values. Such hysteresis was produced by the soil spatial variability rather than the hysteresis of the local parameter values.

Similarly to unsaturated flow, for infiltration there is no single effective conductivity in the general case. However, for very simple conditions, effective parameters may exist. Specifically, for ponded infiltration (i.e. all the water the soil can infiltrate is available at the surface) Rogers (1992) found the geometric mean of the conductivity to be an effective value using either the Green and Ampt (1911) or the Philip's (1957) method. This is based on the assumption of a log-normal spatial distribution of hydraulic conductivity. For non-zero time to ponding, Sivapalan and Wood (1986), using Philip's (1957) equation, showed that effective parameters do not exist, i.e. the point infiltration equation does no longer hold true for spatially variable conductivity.

Overland flow is also a non-linear process so, strictly speaking, no effective parameter exists for the general case. However, Wu *et al.* (1978; 1982) showed that reasonable approximations do exist for certain cases. Wu *et al.* investigated the effect of the spatial variability of roughness on the runoff hydrographs for an experimental watershed facility surfaced with patches or strips of butyl rubber and gravel. Wu *et al.* concluded that an effective roughness is more likely to exist for a low contrast of roughnesses, for a direction of strips normal to the flow direction and for a large number of strips. They also suggested that the approximate effective roughness can be estimated by equating the steady-state surface water storage on the hypothetical uniform watershed to that on the non-uniform watershed. In a similar analysis, Engman (1986) calculated effective values of Manning's n for plots of various surface covers based on observations of the outflow hydrograph. Abrahams *et al.* (1989) measured distribution functions of flow depths and related them to the mean flow depth.

All these studies analysed a single process only. If a number of processes are important, it may be possible to define approximate effective parameters, but their relationship with the underlying distribution is not always clear. Binley *et al.* (1989) analysed, by simulation, three-dimensional saturated–unsaturated flow and surface runoff on a hillslope. For high-conductivity soils, they found effective parameters to reasonably reproduce the hillslope hydrograph, although there was no consistent relationship between the effective values and the moments of the spatial distributions. For low-conductivity soils, characterized by surface flow domination of the runoff hydrograph, single effective parameters were not found to be capable of reproducing both subsurface and surface flow responses. At a much larger scale, Milly and Eagleson (1987) investigated the effect of soil variability on the annual water balance. Using Eagleson's (1978) model they concluded that an equivalent homogeneous soil can be defined for sufficiently small variance of the soil parameters.

The use of effective parameters in microscale equations for representing processes at the macroscale has a number of limitations. These are particularly severe when the dominant processes change with scale (Beven, 1991). Often the dominant processes change from matrix flow to preferential flow when moving to a larger scale. Although it may be possible to find effective values so that the matrix flow equation produces the same output as the preferential flow system, it does so for the wrong reasons and may therefore not be very reliable. A similar example is the changing relative importance of hillslope processes and channel processes with increasing catchment size. Ideally, the equations should be directly derived at the macroscale rather than using effective parameters. However, until adequate relationships are available, effective parameters will still be used.

Disaggregating state variables and inputs

The purpose of disaggregation is, given the average value over a certain domain, to derive the detailed pattern within that domain. Owing to a lack of information for a particular situation, the disaggregation scheme is often based on stochastic approaches. Here, two examples are reviewed. The first refers to disaggregating soil moisture in the space domain and the second refers to disaggregating precipitation in the time domain.

Disaggregating soil moisture may be required for estimating the spatial pattern of the water balance as needed for many forms of land management. The input to the disaggregation procedure is usually a large-scale 'average' soil moisture. This can be the pixel soil moisture based on satellite data, an estimate from a large-scale atmospheric model or an estimate derived from a catchment water balance. One problem with this 'average' soil moisture value is that the type of averaging is often not clearly defined. Nor does it necessarily represent a good estimate for the 'true' average value in a mass balance sense. Although the associated bias is often neglected, there is a substantial research effort underway to address the problem. One example is the First International Satellite Land Surface Climatology Project (ISLSCP) Field Experiment (FIFE) (Sellers *et al.*, 1992). Guerra *et al.* (1993) derived a simple disaggregation scheme for soil moisture and evaporation. The scheme is based on the wetness index and additionally accounts for the effects of spatially variable radiation and vegetation. It is important to note that the wetness index [Equation (4)] along with Equation (5) can be used for both disaggregating and distributing soil moisture. Equation (5) can, in fact, be interpreted as a disaggregation operator (Sivapalan, 1993), yielding the spatial pattern of saturation deficit with the areal average of saturation deficit given. In a similar application, Sivapalan and Viney (1994a,b) disaggregated soil moisture stores from the 39 km^2 Conjurunup catchment in Western Australia to 1–5 km^2 subcatchments, based on topographic and land use characteristics. Sivapalan and

Viney (1994a,b) tested the procedure by comparing simulated and observed runoff for individual subcatchments.

Disaggregating precipitation in time is mainly needed for design purposes. Specifically, disaggregation schemes derive the temporal pattern of precipitation within a storm, with the average intensity and storm duration given. The importance of the time distribution of rainfall for runoff simulations has, for example, been shown by Woolhiser and Goodrich (1988). Mass curves are probably the most widely used concept for disaggregating storms. Mass curves are plots of cumulative storm depths (normalized by total storm depth) versus cumulative time since the beginning of a storm (normalized by the storm duration). The concept is based on the recognition that, for a particular location and a particular season, storms often exhibit similarities in their internal structure despite their different durations and total storm depths. The time distribution depends on the storm type and climate (Huff, 1967). For example, for non-tropical storms in the USA maximum intensities have been shown to occur at about one-third of the storm duration into the storm, whereas for tropical storms they occurred at about one-half of the storm duration (Wenzel, 1982; USDA-SCS, 1986). Mass curves have recently been put into the context of the scaling behaviour of precipitation (Koutsoyiannis and Foufoula-Georgiou, 1993). A similar method has been developed by Pilgrim and Cordery (1975), which is a standard method in Australian design (Pilgrim, 1987). A number of more sophisticated disaggregation schemes (i.e. stochastic rainfall models) have been reported (e.g. Woolhiser and Osborne, 1985) based on the eary work of Pattison (1965) and Grace and Eagleson (1966). A review is given in Foufoula-Georgiou and Georgakakos (1991).

Another application of disaggregating temporal rainfall is to derive the statistical properties of hourly rainfall from those of daily rainfall. The need for such a disaggregation stems from the fact that the number of pluviometers in a given region is always much smaller than that of daily read rain gauges. Relationships between hourly and daily rainfall are often based on multiple regression equations (Canterford *et al.*, 1987).

Up/downscaling in one step

There are a number of methods in use for up/downscaling that do not explicitly estimate a spatial (or temporal) distribution. This means that these methods make some implicit assumption about the distribution and skip the intermediate step in Figure 1. An example is the Thiessen (1911) method for estimating catchment rainfall. It assumes that at any point in the watershed the rainfall is the same as that at the nearest gauge. The catchment rainfall is determined by a linear combination of the station rainfalls and the station weights are derived from a Thiessen polygon network. Another example is depth–area–duration curves. These are relationships that represent the average rainfall over a given area and for a given time interval and are derived by analysing the isohyetal patterns for a particular storm (WMO, 1969; Linsley *et al.*, 1988). Depth–area–duration curves are scaling relationships as they allow the transfer of information (average precipitation) across space scales (catchment sizes) or time-scales (time intervals).

If recurrence intervals or return periods (Gumbel, 1941) are included in the definition of scale, the wide area of hydrological statistics relates to scale relationships. Up/downscaling then refers to the transfer of information (e.g. flood peaks) across time-scales (recurrence intervals).

Linking conceptualizations

Linking conceptualizations across scales can follow either an upward or a downward route (Klemeš, 1983). The upward approach attempts to combine, by mathematical synthesis, the empirical facts and theoretical knowledge available at a lower level of scale into theories capable of predicting processes at a higher level. Eagleson (1972) pioneered this approach in the context of flood frequency analysis. This route has a great appeal because it is theoretically straightforward and appears conceptually clear. Klemeš (1983), however, warned that this clarity can be deceptive and that the approach is severely limited by our incomplete knowledge and the constraints of mathematical tractability (Dooge, 1982; 1986). A well-known example is that of upscaling Darcy's law (a matrix flow assumption), which can be misleading when macropore flow becomes important (White, 1988; Beven, 1991). Klemeš (1983) therefore suggested adopting the downward approach, which strives to find a concept directly at the level of interest (or higher) and then looks for the steps that could have led to it from a lower level. It is clear that the 'depth of inference' (i.e.

the range of scales over which a concept can be inferred) is much more limited in the downward case. Therefore, ideally, the upward and downward search should be combined to form the basis of testable hypotheses (Klemeš, 1983). We will give examples for the two approaches from different areas in hydrology.

The example for the upward approach relates to deriving a macroscale equation for saturated flow in a stochastic framework based on the small perturbation approach (Bakr *et al.*, 1978; Gelhar, 1986). The approach assumes that the local hydraulic conductivity K and local piezometric head H are realizations of a stochastic process and composed of two components

$$K = \bar{K} + k \qquad H = \bar{H} + h \tag{9}$$

where $\bar{K}$ and $\bar{H}$ are the large-scale components and k and h are the local fluctuations. In other words, $\bar{K}$ and $\bar{H}$ are assumed to be smooth functions in space and correspond to mean quantities whereas k and h are realizations of a zero-mean stochastic process. It is clear that the decomposition suggested depends on the scale of the problem examined. For example, what represents the mean at a small-scale laboratory model might be viewed as a fluctuation on a large-scale problem. Combining Equation (9) with the flow equation gives, after a series of assumptions (Bakr *et al.*, 1978), an equation for the mean behaviour. This equation is similar to the local flow equation, but involves additional terms (e.g. representing the covariances of heads and conductivities) which need to be parameterized. This way it gives a new equation, especially if the new additional terms are larger in magnitude than the original terms (Cushman, 1983).

An example of the downward approach has been given by Klemeš (1983). Klemeš related monthly precipitation to monthly runoff in a 39 000 km^2 basin in Canada. Klemeš found a poor relationship and went back to hypothesize about the reasons. He consequently included the effect of evaporation, gravity storage and tension storage in steps. At each step he tested the hypothesis by examining the data and could finally separate the effects of gravity and tension storage. It is important that these results have been arrived at by analysis rather than by postulating them *a priori*. Along similar lines, Gutknecht (1991b) derived a flood routing model for an Austrian basin. Based on a multiple linear storage model, large events were consistently overestimated. Gutknecht (1991b) hypothesized about the reasons and included the effects of inundations for floods higher than a certain stage. He then tested this hypothesis by examining the data and the forecasts were consistently improved. A number of model components were included in steps (in case they turned out to be important), which allowed model complexity to be minimized.

Distributed parameter hydrological models.

Nature of distributed parameter models. Distributed parameter models attempt to quantify the hydrological variability that occurs at a range of scales by subdividing the catchment into a number of units (i.e. subareas). These units may either be so-called hydrological response units (e.g. Leavesley and Stannard, 1990), subcatchments (Sivapalan and Viney, 1994a,b), hillslopes (e.g. Goodrich, 1990), contour-based elements (e.g. Grayson *et al.*, 1992) or, for convenience, square grid elements (e.g. Abbott *et al.*, 1986). In such an approach, processes with a characteristic length scale smaller than the grid/element size are assumed to be represented implicitly (= parameterized), whereas processes with length scales larger than

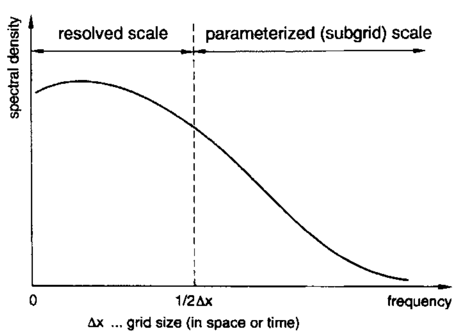

Figure 9. Schematic power spectrum of a hydrological process. Processes with a characteristic scale larger than the grid size are resolved whereas processes with a characteristic scale smaller than the grid size are parameterized as subgrid variability

the grid size are represented explicitly by element to element variations (Smagorinsky, 1974). This is shown in Figure 9, based on a schematic representation of the spectrum of some hydrological process. For example, such a spectrum might be arrived at by, conceptually, sampling infiltration rates along a transect in a catchment and transforming these into the frequency domain. The spectrum in Figure 9 indicates that large-scale processes (low frequency) have more spectral power (= variance) than small-scale processes, which is typical of many hydrological phenomena. It is therefore reasonable to explicitly represent the large-scale processes.

The representation of processes within a unit (i.e. element) involves (a) local (or site) scale descriptions and (b) some assumptions on the variability within the unit. Distributed parameter hydrological models often represent local phenomena (e.g. site-scale infiltration) in considerable detail, while the variability within a unit (i.e. subgrid variability) is often neglected. To drive the models for each unit, input variables (e.g. precipitation) need to be estimated for each element. This involves some sort of interpolation between observations. Conversely, the response from individual elements is coupled by routing routines. Distributed parameter catchment models have been reviewed by Goodrich and Woolhiser (1991) and Moore *et al.* (1991) and recent contributions are given in Beven (1992) and Rosso (1994).

Many arguments have been put forward in favour of distributed models (e.g. Beven *et al.*, 1980). Although distributed models never lived up to their expectations in terms of their performance in predicting runoff (Loague, 1990; Obled, 1990), this has been early recognized by some workers (Freeze and Harlan, 1969). Bergström (1991) presented an excellent discussion on modelling philosophy in hydrology in general and model complexity in particular. Clearly, the optimum model complexity depends on the nature of a specific problem. There is a wide range of problems where bulk models cannot do the job and distributed parameter models are called for. These include those where spatially distributed estimates of runoff are required and/or a high degree of process understanding is needed (Bergström, 1991).

Unfortunately, distributed models invariably suffer from a number of limitations (Beven, 1989; Grayson *et al.*, 1992; 1993; Kirnbauer *et al.*, 1994). These include (a) the extreme heterogeneity of catchments, which makes it difficult to accurately define element to element variations and subgrid variability and (b) the large number of model parameters, which makes model calibration and evaluation very difficult.

Both aspects are scale-related issues. The quantification of element to element variations has been examined earlier in this paper (distributing information). The representation of subgrid variability and model evaluation are now being discussed.

Subgrid variability. There are three approaches for quantifying the variability of hydrological processes within a computational element (grid cell).

The first approach assumes that the parameters and processes are uniform within each element and that the local (small-scale) descriptions apply to the whole element. The local parameters are then replaced by effective parameters (see section on aggregating model parameters). However, effective parameters do not always exist, particularly when the processes are non-linear.

The second approach uses distribution functions rather than single values. This has advantages for non-linear systems, but complicates identifiability significantly. Goodrich (1990) used distribution functions of saturated hydraulic conductivity for representing variability on a hillslope as a part of a catchment model. A hillslope was decomposed into a number of non-interacting strips in parallel and the values for each strip were assigned according to a log-normal distribution. In a similar fashion, Famiglietti (1992) modelled the spatial variability of evaporation by a distribution function based on the wetness index [Equation (4)]. The distribution of the index was discretized into a number of intervals and the local model was applied to each interval. These and similar techniques are clearly related to methods of distributing information.

The third approach parameterizes subgrid variability without explicitly resorting to the local equations. An excellent example is the parameterization of overland flow in rills suggested by Moore and Burch (1986). In this parameterization individual rills are not modelled explicitly, but rather the lumped effect of a larger number of rills is represented by a power law of the form

$$R = \xi A^m \tag{10}$$

where R is the hydraulic radius, A is the cross-sectional area and ξ and m are parameters. It is important to note that the information on the detailed geometry of the rills is replaced by only two parameters. This is reminiscent of Darcy's law, where the detailed information on pore geometry is replaced by one lumped parameter (i.e. hydraulic conductivity). The exponent m is unity for sheet flow and 0·5 for trapezoidal or parabolic geometries. For most natural surfaces m lies between these limits (Foster *et al.*, 1984; Parsons *et al.*, 1990; Willgoose and Riley, 1993). The representation of overland flow also illustrates well the partitioning of topographic variability into resolved and parameterized variability (Figure 9) as used in some models (Grayson *et al.*, in press). The resolved variability (i.e. element to element variability) is represented by the shapes, slopes and aspects of individual elements and their relative configuration. The parameterized (i.e. subgrid) variability is represented by the lumped Equation (10). As long as these two representations are consistent, the element size should have little effect on the results. However, this is not always so. Willgoose and Riley (1993), for example, showed that the parameters of Equation (10) can vary significantly with catchment size (see also Willgoose and Kuczera, 1995).

If there is a minimum in the power spectrum of the process to be modelled (Figure 9), i.e. a spectral gap, the quantification of subgrid variability becomes easier. Specifically, the same parameterization can be used for a range of element sizes as long as the element size falls into the spectral gap. This is discussed in more detail in Blöschl *et al.* (1995).

Model evaluation. The difficulty in evaluating and calibrating distributed models on the basis of catchment runoff (e.g. Blöschl *et al.*, 1994) lead Bathurst and Cooley (in press) to remark 'A complicating feature of multiple parameter models is the possibility that apparently equally satisfactory simulations can be achieved with different combinations of physically realistic parameter values, the change in one parameter being compensated for by a change in another.' In other words, runoff, being an integrated value, cannot easily identify the high-frequency component of spatial parameter distributions. One way to address this problem is to use subcatchment runoff for cross-checking (e.g. Sivapalan and Viney, 1994a,b). However, in practice the number of subcatchments for which data are available is often very limited. Also, groundwater levels and tensiometer readings within the catchment have been used (Koide and Wheater, 1992), but these are invariably point values and not always representative. As an alternative it has been suggested to use spatial *patterns* of state variables to assess the accuracy of models within a catchment (Blöschl *et al.*, 1994). Patterns have a very high *space* resolution as opposed to a very high *time* resolution of time series such as hydrographs. One notable example is Moore and Grayson (1991), who evaluated a distributed parameter model on the basis of saturation patterns (i.e. zones of surface saturation). Their analyses were based on a $2\,m^2$ laboratory catchment. Moore and Grayson (1991) demonstrated that accurate runoff simulations at the catchment outlet do not necessarily imply accurate simulations of distributed catchment response. In a similar fashion, Blöschl *et al.* (1991) used snow cover depletion patterns to evaluate a distributed hydrological model in a $10\,km^2$ Alpine catchment. They showed that the patterns allowed the assessment of the accuracy of individual model components (such as radiative exchange) and discrimination between alternative model assumptions (Blöschl *et al.*, 1994). It is clear that distributed model evaluation is the key to progress in distributed modelling. However, significantly more data are needed in the future. This particularly refers to reliable measurements of soil moisture patterns, which are ideal for the evaluation of distributed models.

DIMENSIONAL ANALYSIS AND SIMILARITY

Dimensional techniques

Dimensional techniques are powerful for dealing with complex physical problems, as they can potentially describe these systems by very simple relationships. Dimensional techniques have been responsible for some of the great advances in hydraulics (Fischer *et al.*, 1979) and yet have only been used sparingly in catchment hydrology (Dooge, 1986).

Dimensional techniques can be useful in a number of ways. They can be used to define similarity relationships (between two catchments or between a catchment and a scale model) in the same fashion as the Froude and Reynolds numbers have been used in hydraulics. Further, they can be used to establish relationships that are valid over a wide range of scales in the same fashion as the Moody diagram (e.g. Chow *et al.*, 1988) is being used for pipe network design. Also, they are powerful for data reduction. This makes them appealing for regional flood frequency analysis and regionalized loss estimation (NRC, 1988; Wood *et al.*, 1990; Beven, 1991) as they allow for classifying data into zones of similar behaviour. Finally, they have potential for characterizing the dominant hydrological mechanisms operating on a specific catchment.

The foundation of all dimensional techniques rests in the concept of similarity. Similarity exists between two systems whenever the characteristics of one system can be related to the corresponding characteristics of another system by a simple conversion factor, called the scale factor (Langhaar, 1951). Three types of similarity are possible in physical systems: geometric, kinematic or dynamic (Tillotson and Nielsen, 1984). Geometric similarity relates to the size relationship between two systems. When two catchments of different size have similar shapes they are considered to be geometrically similar. Kinematic similarity applies when the ratios of velocities in the two systems are constant multiples of each other. When the instantaneous unit hydrographs of two catchments are related by a constant scale factor, they can be said to be kinematically similar. One example of kinematic similarity relationships in catchment hydrology is the geomorphologic unit hydrograph (Rodríguez-Iturbe and Valdés, 1979). However, it is well known that small catchments behave differently from large catchments due to the effects of spatial heterogeneity and changing processes, among other things, and dynamic similarity may need to be invoked to describe these.

In hydraulics, dynamic similarity is said to exist when the forces in one system are a constant multiple of the equivalent forces in the other. For example, suppose two flow systems depend on inertial and viscous forces only. They are dynamically similar if the ratio of viscous to inertial forces (the Reynolds number) is identical between the two systems. In general, dynamic similarity reflects a balance of two dynamic quantities. This can be a balance of forces, a balance of energy, of mass, of momentum, etc. In catchment hydrology the term dynamic similarity may be ambiguous in meaning, as it is not perfectly clear whether we would consider ratios that we know clearly capture the 'dynamics' of catchment responses as truly representing dynamic similarity. Examples are the ratio of surface to subsurface runoff or the ratio of hillslope response times to stream network response times. Strictly speaking, only kinematic quantities are involved in both examples. Clearly, more careful thought needs to be given to a definition of similarity appropriate to catchment hydrology.

All methods used to determine scale factors can be grouped into three main classes: dimensional analysis, similarity analysis and functional normalization (Tillotson and Nielsen, 1984). Dimensional anlaysis and similarity analysis allow the definition of geometric, kinematic or dynamic similarity, whereas functional normalization does not. In this section we give brief definitions of these techniques and examples of their applications in hydrological modelling and flood estimation.

Dimensional analysis. Dimensional analysis is a simplification process by which the number of dimensional quantities used to describe a system is reduced to a fewer number of non-dimensional quantities called π terms. The number of non-dimensional π terms needed to completely define a system is given by the Buckingham pi theorem (Buckingham, 1914). Note that dimensional analysis can be carried out even when the laws or equations governing the physical system are unknown. In fact, it is a particularly useful technique in such instances.

The systematic procedure by which dimensional analysis should be carried out for any selected physical problem is presented in many standard texts (Fischer *et al.*, 1979; Stull, 1988). This involves two main steps: (a) select the variables relevant to the problem and organize the variables into non-dimensional groups; and (b) perform an experiment, or gather the relevant data from previous experiments, to determine the values of the non-dimensional groups and determine relationships between the non-dimensional groups based on the empirical data. The first step can potentially give dimensionless numbers (such as the Froude number), which could be used to design scale experiments. The second step, in case consistent relationships can be found, is even more useful as it allows the establishment of 'universal' relationships such as those represented in the Moody diagram.

Similarity analysis. As in dimensional analysis, similarity analysis seeks to (a) organize variables into non-dimensional groups and to (b) determine relationships between the groups (Hellums and Churchill, 1961; Tillotson and Nielsen, 1984). The second step, i.e. determining relationships between the groups, follows along the lines of dimensional analysis and can be carried out by field (or laboratory) experiments and/or numerical simulations. However, the first step, i.e. the organization of variables into non-dimensional groups, is different. Specifically, unlike dimensional analysis, similarity analysis requires that physical laws or equations governing the system of interest are known. The idea in similarity analysis is to rewrite the partial differential equations in non-dimensional form, which gives the dimensionless groups. Similarity analysis is more powerful than dimensional analysis (Miller, 1990). For example, it can handle more than one parameter with a given unit (e.g. more then one quantity of unit length) and it can also handle dimensionless quantities, which is not the case with dimensional analysis. A classical example of similarity analysis in soil physics is the work of Miller and Miller (1956).

Functional normalization. Unlike the two physically based dimensional techniques discussed earlier, functional normalization (Tillotson and Nielsen, 1984) is an empirical method. It begins with a set of empirical relationships between two or more variables such as soil moisture characteristic relations. The objective then is to coalesce all such empirical relationships in the set into one, or a few, reference curve(s) that describe the set as a whole. Functional normalization is therefore a simplification process which is empirical and does not necessarily have any physical justification.

Hydrological similarity of catchment responses

Dimensional analysis of catchment form and dynamics. Strahler (1964) suggested that catchments with common underlying geology, climate and lithology contain a high degree of geometrical similarity. 'Horton's laws' of network morphology (Horton, 1945; Strahler, 1957) (which are discussed later in this paper), describe a form of geometrical similarity. Note that these are not physical laws, but rather empirical laws and the process by which they are derived is an example of functional normalization.

Strahler (1964) made the first attempt at applying dimensional analysis to examine the similarity of catchment form based on its dynamics. Through application of the Buckingham pi theorem, Strahler suggested that the normalized drainage density of catchments (denoted by HD) can be related to three physically based non-dimensional groupings, or similarity parameters

$$HD = \Pi(Q_r K,\ Q_r \rho H/\mu,\ Q_r^2/Hg) \tag{11}$$

where Q_r is runoff intensity (rainfall intensity minus infiltration capacity), K is an erosion proportionality factor defined as the ratio of erosion intensity to erosion force (a measure of erodibility of the ground surface), D is the drainage density, H is the vertical relief representing the potential energy of the system, ρ is the density of water, μ is viscosity and g is the acceleration due to gravity. $Q_r K$ is called the Horton number, which expresses the relative intensity of erosion processes within the catchment. $Q_r \rho H/\mu$ is a form of the Reynolds number in which Q_r takes the place of the velocity and H is the characteristic length. Finally, Q_r^2/Hg is a form of the Froude number. The reduction of the seven dimensional variables to four non-dimensional groups, shown in Equation (11), 'focuses attention upon dynamic relationships, simplifies the design of controlled empirical observations, and establishes conditions essential to the validity of comparisons of models with prototypes' (Strahler, 1964: 4–71). However, the actual functional form of Π in Equation (11) has not yet been determined. Also, it is not clear whether the non-dimensional terms are measurable in practice.

Similarity analysis of runoff generation. As runoff generation is a complex phenomenon, involving tremendous spatial heterogeneity, it has been difficult to generate universally applicable concepts and theories to describe runoff generation. Sivapalan *et al.* (1987) applied similarity analysis to gain insights into the processes of runoff generation. The starting point was a physically based model of runoff generation, based on TOPMODEL concepts (Beven and Kirkby, 1979), which included both infiltration excess and saturation runoff generation.

The first step in the similarity analysis involved recasting the model equations into non-dimensional forms. All variables in the model equations were first transformed to non-dimensional form by dividing each by a fixed arbitrary reference quantity of the same dimension. These were then substituted into the model equations. The resulting equations were then rearranged to their non-dimensional forms in such a way that the coefficient of the leading term of every equation, often the left-hand side of the equation, was equal to unity. The similarity analysis along these lines by Sivapalan *et al.* (1987) resulted in the identification of five non-dimensional similarity parameters and three auxiliary variables. It was believed that these parameters embody the interrelationships of topography, soil and rainfall that lead to similar catchment responses. However, the identification of these non-dimensional groupings is only the first step in similarity analysis. The second, and perhaps even more important, step is the derivation of relationships among the similarity parameters, i.e. similarity relationships.

Larsen *et al.* (1994) carried out an extension of the similarity analysis of Sivapalan *et al.* (1987), which resulted in the derivation of a number of similarity relationships. Their study was carried out on a number of small agricultural catchments in the eastern wheat belt of Western Australia. The study concentrated on the relative dominance of two alternative mechanisms of runoff generation in the region: infiltration excess and saturation excess. To quantify this, Larsen *et al.* (1994) defined a runoff measure, denoted by R, as the ratio of the volume of simulated saturation excess runoff to the volume of simulated combined runoff, integrated over a hypothetical year. If $R = 1$, then all runoff is of the saturation excess type, whereas if $R = 0$, then all runoff is of the infiltration excess type. According to Larsen *et al.*'s similarity theory, the dependence of R on topography, soils and rainfall, and their spatial variability, can be expressed in terms of five non-dimensional similarity parameters and two auxiliary variables, as follows:

$$R = \Psi\{p^*, t_d^*; f^*, K_0^*, B^*, C_{V,K_0}, C_{V,\ln(a/\tan\beta)}\} \tag{12}$$

where p^* and t_d^* are auxiliary climatic parameters relating to the mean rainfall intensity and mean storm duration, and f^*, K_0^*, B^*, C_{V,K_0}, $C_{V,\ln(a/\tan\beta)}$ are the non-dimensional similarity parameters (f^* is a measure that is inversely related to the storage capacity of the soil profile; K_0^* is a measure of the infiltration capacity of the soil; B^* is related to the pore size distribution; C_{V,K_0} and $C_{V,\ln(a/\tan\beta)}$ are the coefficients of variation (in space) of K_0^* and the wetness index, respectively). The purpose of similarity analysis is to derive Equation (12) based on a combination of field or laboratory data and simulations.

By means of sensitivity analysis with their model, and analysis of field data from seven small agricultural catchments in the eastern wheat belt, Larsen *et al.* (1994) showed that only the first four parameters in Equation (12) were important for predicting R (i.e. predicting which runoff generation process dominated in these catchments). Larsen *et al.* (1994) then carried out extensive simulations, with artificial catchments derived by varying the four critical similarity parameters. Each simulation resulted in a value of R which could then be used to infer the complex similarity relationship among p^*, t_d^*, f^* and K_0^*. For

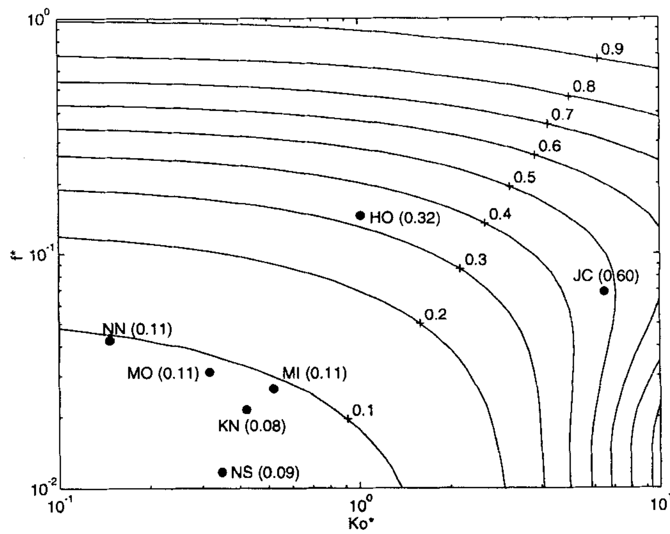

Figure 10. Contours of R values (ratio of simulated saturation excess runoff volume to simulated total runoff volume) as a function of K_0^* (a measure of infiltration capacity) and f^* (a measure that is inversely related to the storage capacity of the soil). The points represent actual catchments. From Larsen *et al.* (1994)

verification, Larsen *et al.* (1994) also estimated the values of R for each of the seven experimental (actual) catchments.

Larsen *et al.* (1994) presented their results in the form of contour plots of the variation of R in K_0^*–f^* space and $p^*-t_d^*$ space. Figure 10 is an example of the similarity relationships so obtained, and presents the variation of R in K_0^*–f^* space. The R values for the seven actual catchments are presented within brackets alongside their positions on the diagram. Both the parameters f^* and K_0^* have physical meanings which are directly related to the two runoff generation mechanisms. In Figure 10, the bottom left-hand corner, for example, relates to deep soils (small f^*) of low infiltration capacity (small K_0^*), hence runoff is mainly of the infiltration excess type. This is clearly consistent with physical reasoning. The contours in Figure 10 are similarity relationships between the parameters f^* and K_0^* for catchments in the eastern wheat belt region. The R values for the actual catchments are consistent with the values predicted by the similarity theory (contour lines). This demonstrates that catchment responses can indeed be scaled using the similarity parameters f^* and K_0^*.

Robinson and Sivapalan (1995) have further extended the work of Larsen *et al.* (1994). They investigated the regionalization, based on the same concepts of similarity, of a lumped catchment-scale runoff prediction equation which included both infiltration excess and saturation excess mechanisms of runoff generation. One of the advantages of this approach is that the regionalization is based on runoff generating mechanisms and field soil properties that, at least in principle, can be measured in the field. In reality, such field estimation is not a trivial exercise.

Regionalization of flood frequency.

Empirical flood frequency analysis: functional normalization

One of the practical motivations for the investigation of hydrological similarity is the widely perceived need to improve the physical basis for the grouping of catchments for regional flood frequency analysis. Ideally, regionalization of flood frequency should be based on the actual flooding mechanisms operating on catchments within the region. However, in current practice, such regionalizations are carried out mainly on empirical grounds with only a few exceptions (e.g. Kölla, 1987). The empirical procedures used for this exercise belong to the category of functional normalization. This is because they are aimed at collapsing a large number of empirical relationships (flood magnitude against probability of exceedance for different catchment types and sizes) into one or a few regional relationships(s).

Empirical regionalization of flood frequency involves two steps: forming catchment groupings, and coalescing all of the flood frequency curves within each group into one characteristic regional curve. The latter is usually done by scaling the flood peaks by the mean annual flood. This procedure is known as the index flood method (Flood Studies Report, 1975; NRC, 1988).

Very sophisticated techniques are presently available for grouping catchments based on objective statistical criteria (Wallis, 1988; Wiltshire, 1985). An excellent example of the application of these techniques for regional flood frequency analysis is provided by work carried out in New Zealand, as summarized in McKerchar and Pearson (1989) and Pearson (1993).

Model-based derived flood frequency: similarity analysis

The use of similarity analysis in combination with the derived flood frequency approach was pioneered by Eagleson (1972). Considerable work has been carried out since by Wood and Hebson (1986) and Hebson and Wood (1986), who applied similarity analysis to gain insights into the effects of catchment size (in relation to storm size) on the flood frequency curve. However, much of this work was concerned with runoff routing rather than runoff generation.

Sivapalan *et al.* (1990) extended the derived flood frequency work of Wood and Hebson (1986) and Hebson and Wood (1986) by explicitly considering aspects of runoff generation. Their derived flood frequency model combined the runoff generation model of Sivapalan *et al.* (1987) and a generalized geomorphologic unit hydrograph (based on Horton order ratios) to handle partial area runoff generation. The combined model was rewritten into a non-dimensional form by adopting the methods of similarity analysis used by Sivapalan *et al.* (1987). Using this model Sivapalan *et al.* (1990) carried out extensive simulations to investigate the sensitivity of the flood frequency characteristics to the similarity parameters.

An important conclusion of the work of Sivapalan *et al.* (1990) was that there is indeed a connection

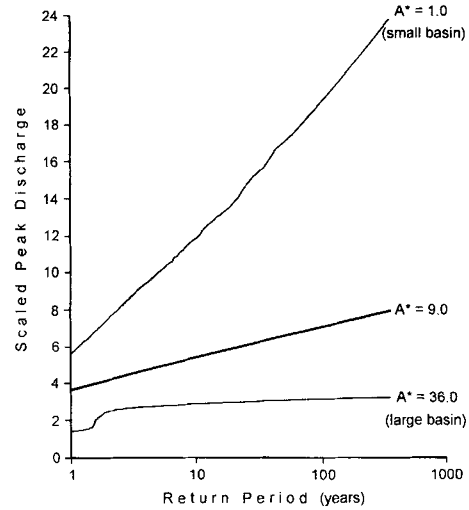

Figure 11. Sensitivity of the flood frequency distribution to the scaled catchment area A^* from Sivapalan *et al.* (1990)

between the shape of the flood frequency curve and the relative dominance of the mechanisms of runoff generation. In view of the groupings of the eastern wheat belt catchments obtained by Larsen *et al.* (1994) in terms of the dominant mechanisms of runoff generation, i.e. the ratio R, any possible connection between the empirical and physically based groupings of the catchments merits very careful examination. Hebson and Wood (1986) and Sivapalan *et al.* (1990) also showed that the ratio of the spatial correlation length of the storms to the catchment size was critical to the shape of the flood frequency curve. This is illustrated by Figure 11, taken from Sivapalan *et al.* (1990), which indicates that the shape of the flood frequency curve changes with catchment scale, becoming steeper with decreasing catchment size (relative to storm size). Current work by Gupta and Dawdy (1995; 1994) suggests that these conclusions are consistent with a theoretical framework of regional flood frequency based on multi-scaling arguments. Further investigation of these connections is worthwhile as they may have wide-ranging practical significance.

Self-similarity and fractals

Fractals, like other similarity concepts, can potentially describe complex phenomena by a minimum of parameters, which makes them an appealing concept. Fractals are based on the recognition that variability exists at a range of scales. In fact, fractals allow the relationship of variability between different scales to be quantified. The main use of fractal concepts in hydrology is in determining this relationship. This can be the quantification of rainfall variability as a function of time- (or space) scale or the quantification of soil parameters as a function of space scale. Once the relationship is derived for certain conditions it is possible to extrapolate the rainfall (or soil parameter) variability to a larger or smaller time- (or space) scale. Similar properties of fractals are used in analyses of the structure of stream networks. One specific application is the definition of a critical scale at which the natural phenomena (e.g. topography) deviate from the 'ideal' fractal relationship. This seems to have potential to discriminate between hillslope and channel processes (Tarboton *et al.*, 1991; Montgomery and Foufoula-Georgiou, 1993). It has also been attempted to use fractals to *classify* landscapes (Klinkenberg and Goodchild, 1992), soils and other variables.

Fractals are particularly convincing by their visual appearance (such as simulated clouds), which often resembles natural phenomena (e.g. Lovejoy and Schertzer, 1985). However, this may be deceptive as fractals, generally, are non-physical approaches and care must be exercised in their interpretation. It will be a challenge to investigate to which degree fractals can be compatible with physical reasoning (Gupta and Dawdy, 1995; 1994).

What are fractals? Simple fractals are sets with certain properties independent of scale (Voss, 1988; Cutler, 1993). Fractals can be either deterministic or random. In the case of deterministic (geometric) fractals, the property independent of scale is the geometric shape of the set. In the case of random (statistical) fractals, these properties independent of scale are certain statistical properties of that set (Feder, 1988). The selection of the statistical properties that are required to be independent of scale depends on whether an ordered set or an unordered set is considered. Ordered sets are sets where a quantity is associated with a location (in space or time), whereas unordered sets are simply collection of objects. Specifically, unordered sets are said to be simple random fractals if their probability density function (pdf) is independent of scale (e.g. area–perimeter relationships; Lovejoy, 1982; Rodríguez-Iturbe *et al.*, 1992a), whereas ordered sets (i.e. mathematical processes) are said to be simple random fractals if properties such as the pdf of their *increments* are independent of scale (see Mandelbrot and Van Ness, 1968; Klinkenberg and Goodchild, 1992). We will detail the requirements for ordered sets to be simple random fractals as these are most commonly used in hydrology. Usually, one of the following three relationships is examined.

(a) Scaling of the increments. In the first description, the *pdf of the increments* of a process is required to be independent of scale (Mandelbrot and Van Ness, 1968; Lovejoy and Schertzer, 1989). Such a pdf is called, equivalently, scaling, self-similar or hyperbolic. The condition of independence of scale is satisfied by a power law

$$\{Z(x+h) - Z(x)\} \stackrel{d}{=} \lambda^{-H}\{[Z(x+\lambda h) - Z(x)]\} \tag{13}$$

where h is the spacing (lag) between two points x and $x+h$; λ is the scaling parameter; H is a scaling exponent; $Z(x+h) - Z(x)$ is the increment (fluctuation) of the property for a given spacing h and point x and the equality $\{\cdot\} \stackrel{d}{=} \{\cdot\}$ indicates equality of pdfs. Equation (13) compares two pdfs. One is derived from a set of increments with a small spacing. For example, this may be a time series of rainfall Z with a spacing (time step) h of, say, five minutes. The other pdf is derived from a set of increments with a larger spacing (λh), which might be a time series of rainfall with hourly time steps. In practical applications, the latter is often arrived at by resampling the former. The ratio of the spacings is the scaling parameter λ ($\lambda = 12$ in the above example).

(b) Semivariogram. In the second description, the *variance of the increments* of a process (i.e. the semivariogram) is required to be independent of scale (Mandelbrot and Van Ness, 1968; Mandelbrot, 1977)

$$\gamma(h) = \lambda^{-2H}\gamma(\lambda h) \tag{14a}$$

and

$$\gamma(h) = 0{\cdot}5E\{[Z(x+h) - Z(x)]^2\} \tag{14b}$$

where $\gamma(h)$ is the semivariogram function and $E\{\ \}$ refers to the expectation (Oliver and Webster, 1986). Clearly, Equation (14) is a relaxation of Equation (13) as only variance is required to scale rather than the complete pdf.

(c) Power spectrum. In the third description the *power spectral density* S of a process is required to be independent of scale (Mandelbrot and Van Ness, 1968; Voss, 1985)

$$S(\lambda f) = \lambda^{-\beta}S(f) \tag{15}$$

where f is the frequency and β is a scaling exponent. Equations (14) and (15) are different representations of the same information. 'Roughly speaking', a power law in both Equations (14) and (15) is consistent (Carlson, 1986: 162ff; Voss, 1988) and Equation (14) is equivalent to Equation (15) with a linear relationship between β and H (e.g. Voss, 1988). Strictly speaking, however, the consistency is not very clear and the relationship between H and β appears to be non-linear (Gallant *et al.*, 1994). Part of the problem is that the power spectrum Equation (15) is undefined as fractals are non-stationary.

The power laws Equations (13), (14) and (15) appear as straight lines in double logarithmic plots. The goodness of fit of data to that straight line is generally used as the one criterion to judge if a given data set may be approximated by a fractal or not (Feder, 1988; Tarboton *et al.*, 1988). However, it is notoriously easy to fit double logarithmic relationships to data (Lewis, 1995). Also, the fractal approximation is often limited by a lower and an upper cutoff (Feder, 1988).

There are a variety of other techniques that are used to determine if a set is a simple random fractal or not (Gallant *et al.*, 1994) and the value of the scaling exponent. These techniques include the divider method (Mandelbrot, 1967) and the rescaled range analysis (Mandelbrot and Wallis, 1969).

Using various techniques, a substantial amount of geophysical data (including rainfall and streamflow) has been examined (e.g. Hurst, 1951; Mandelbrot and Wallis, 1968; 1969; Lettenmaier and Burges, 1977; Lovejoy, 1982; Skoda, 1987; Klinkenberg and Goodchild, 1992). Most analyses showed strong evidence for values of the scaling exponent H [in Equation (13) or (14)] to be greater than 0·5 (typically between 0·7 and 0·8). For a process in which any event is dependent only on the preceding event (a Markov process), the exponent H is 0·5, whereas $H > 0{\cdot}5$ indicates long-term persistence (the Hurst phenomenon). The existence of long-term persistence would question traditional statistical methods in hydrology, particularly those related to design. However, it is not yet clear how robust the methods used in these analyses really are. Specifically, work by Mesa and Poveda (1993) questioned their reliability. Based on a so-called 'GEOS' diagram which they deemed more reliable, Mesa and Poveda found no evidence that H is significantly different from 0·5. More work seems to be needed on that issue.

The preceding section dealt with simple random fractals. Simple fractals have a number of drawbacks, particularly for rainfall modelling (Lovejoy and Schertzer, 1989). Firstly, they involve only two parameters (slope and intercept of the straight line in the double logarithmic plot) and are thus very special. Secondly, they cannot cope with zero rainfall and require thresholding of the fractal. One way of simultaneously overcoming both drawbacks is to use 'multifractals' (multiple scaling).

Multifractals are more general than simple fractals. They consist of intertwined fractal subsets with different scaling exponents (Feder, 1988). Also, they exhibit a very singular limiting behaviour with values everywhere almost surely zero, and all of the mass is concentrated into singularities of various orders. Applications in the context of atmospheric turbulence, rainfall and streamflow have been suggested by Oboukhov (1962), Gupta and Waymire (1990; 1993), Lovejoy and Schertzer (1991), Sreenivasan (1991) and Gupta and Dawdy (1994, 1995). Multifractals involve *multiplicative* modulation of the small scales by the large, as opposed to *adding* random elements at different scales (as is the case for simple fractals). Also, multifractals are defined for *measures*, whereas simple fractals are defined for *sets* (Falconer, 1990). One convenient model of multifractals is the binomial multiplicative cascade (e.g. Gupta and Waymire, 1993). Multifractals can be characterized by their singularity spectrum $f(\alpha)$ (Halsey *et al.*, 1986; Ijjasz-Vasquez *et al.*, 1992), which describes the scaling behaviour of the singularities of a process rather than that of the process itself. Ijjasz-Vasquez *et al.* (1992) argued that the singularity spectrum is 'useful as a first step toward the development of multiplicative cascade models of energy dissipation similar to models implemented in studies of turbulence (Menevau and Sreenivasan, 1987; Chhabra *et al.*, 1989; Menevau *et al.*, 1990).' However, Chhabra *et al.* (1989) indicated that extracting underlying multiplicative processes from the singularity spectrum may be very difficult. Wavelets (e.g. Daubechies, 1992) may have some potential (Tarboton, pers. comm.). It will be a challenge for future research to adequately address this identification problem. It will also be a challenge, once the underlying process is known for a given catchment, to translate it into information of practical relevance (Smith, 1992; Gupta and Dawdy, 1994, 1995).

STREAM NETWORK ANALYSIS

Importance of stream network analysis

Stream network analysis is one area in which dimensional techniques and similarity are particularly important and widely used. This is because dimensional techniques and similarity concepts are particularly powerful for dealing, in a simple way, with a range of complex processes such as those involved in forming stream channels. The original motivation of characterizing stream networks for hydrological applications was the hope that synthetic unit hydrographs can be derived for ungauged catchments based on properties of the stream network. Although this has never been successful for practical applications, it has been useful in understanding the scaling behaviour of catchment response. In other words, stream network analysis provides investigative rather than predictive tools. A further potential and future goal is that the spatial variability of soils might be inferred from stream networks. This is based on the rationale that the processes involved in the formation of stream networks and soils are coupled by a number of feedback mechanisms (Beven *et al.*, 1988; Gupta *et al.*, 1986b).

To describe a network completely, two pieces of information are needed. These are (a) the topology (i.e. map view and elevation) and (b) the cross-sectional geometry and hydraulic properties. Information on both aspects is needed for models such as the geomorphologic unit hydrograph or the metachannel.

Quantitative description of the topology of stream networks

Early work on the quantification of stream networks relied on an ordering system devised by Horton (1945) and later improved by Strahler (1957), which highlighted the natural convergence inherent to stream networks. Following the analysis of a large number of stream networks, Horton (1945) and Schumm (1956) presented a set of empirical relationships, known as *Horton's laws*, which were based on this ordering system. They found that, in a region of uniform geology and climate, the ratios of number of streams, length of streams, area of streams and slopes of streams between successive orders are approximately constant, i.e. independent of order. These constants are generally known as the Horton order ratios, and are given by

$$\text{Bifurcation ratio} \quad R_B = N_{w-1}/N_w \tag{16a}$$

$$\text{Length ratio} \quad R_L = L_w/L_{w-1} \tag{16b}$$

$$\text{Area ratio} \quad R_A = A_w/A_{w-1} \tag{16c}$$

$$\text{Slope ratio} \quad R_S = S_{w-1}/S_w \tag{16d}$$

where N_w is the number of streams of order w, L_w is the mean length of streams of order w, A_w is the mean area contributing to the streams of order w and S_w is the mean slope of streams of order w.

The Horton order ratios [Equations (16)] are dimensionless parameters that are often thought to collectively represent scaling relationships for catchments of different sizes. However, it has been shown (Kirchner, 1993) that almost all possible networks have Horton order ratios similar to those derived for natural catchments. This is considered by Kirchner to imply that Horton's 'laws' are artefacts of stream ordering techniques and as such express little or nothing as regards either the uniqueness of individual networks or indeed the scaling between catchments of different sizes with their attendant networks. Moreover, the scaling of network properties using Horton order ratios becomes even more difficult in regions of changing or non-uniform lithology (the erodibility and infiltration capacity of soils), geology (structural forms) and climate (Horton, 1945; Chorley *et al.*, 1984).

Several other quantitative measures of stream networks have also been proposed, which do not assume the Strahler ordering. Rather, they express the cumulative properties of the network geometry in terms of the flow distance s, or altitude h, from the outlet, in effect collapsing the network into a single 'effective' channel. Some of the more commonly used functions based on the flow distance s or altitude h are: (1) the width function, $N(s)$, which expresses the number of links occurring at a flow distance s from the outlet (Lee and Delleur, 1976; Kirkby, 1976; Mesa and Mifflin, 1986); (2) the distance–area function, $a(s)$, which describes the local area draining per unit length of channel at a flow distance s from the outlet (Snell and Sivapalan, 1994a,b); (3) the link concentration function, $N(h)$, which expresses the number of links occurring at an altitude h above the outlet (Gupta *et al.*, 1986b); and (4) the hypsometric curve (Strahler, 1957), denoted by $A(h)$, describing the cumulative area of the catchment above a given altitude h.

In the past, the methods for estimating the Horton order ratios, and other geomorphological functions, relied heavily on topographic maps and used manual procedures, which are time consuming. With the advent of digital elevation models (DEMs), there has been a proliferation of automatic procedures for processing DEMs, extracting channel networks and estimating the various geomorphological functions and parameters (Band, 1986; Jenson and Domingue, 1988; Band and Moore, 1995). All of these methods define the streams on a DEM as consisting of all points with accumulated drainage area above some threshold value, called a 'support area'.

The choice of support area is often made in a subjective manner as insufficient guidance is available based on physical considerations. There is considerable debate as to whether the use of a support area is the best method to define channel initiation, considering the spatial non-uniformity of geology, soils and rainfall over large geographical regions. As yet, the problem is far from resolved and work is continuing, using both theoretical and field studies, to find the most objective way to define the initiation of stream channels (Tarboton *et al.*, 1991; Dietrich and Dunne, 1993). For example, Montgomery and Foufoula-Georgiou (1993) have shown that a constant critical support area, the method most commonly used at present, is more appropriate for depicting the hillslope/valley transition than for identifying channel heads and suggest that a slope-dependent (i.e. non-constant) critical support area is both theoretically and empirically more appropriate for defining the extent of channel networks. Resolution of this problem is important as it is becoming clear that geomorphological attributes such as order ratios and the width and area functions are heavily dependent on the support area used to extract stream networks from DEMs (Lee and Delleur, 1976; Helmlinger *et al.*, 1993; Snell and Sivapalan, 1994b; Gyasi-Agyei *et al.*, 1995).

Some recent work has attempted to characterize stream network morphology in terms of fractals and multifractals as a way to better quantify the scaling in geomorphological attributes (Tarboton *et al.*, 1988; 1989; Rosso *et al.*, 1991). In particular, most of this work was geared towards the estimation of fractal exponents and to the study of how these exponents can be related to geomorphological laws such as the Horton's laws. Much work is now in progress to build models of hydrological response based on a fractal or multi-fractal characterization of stream networks (Marani *et al.*, 1991; Rinaldo *et al.*, 1991). Although these methods are mathematically sophisticated, it is unlikely that significant progress can be achieved without accounting for the effects of geology, lithology and climate. For example, geology is clearly a dominant control of network structure (e.g. von Bandat, 1962; Drury, 1987). This is probably best known for drainage density (the ratio of total stream length to basin area). Drainage densities can vary more than three orders of magnitude between, say, limestone strata in Germany ($< 0{\cdot}28\,\mathrm{km/km^2}$; Kern, 1994) and badlands developed on clay in New Jersey (about $700\,\mathrm{km/km^2}$; Strahler, 1964). It is clear that such differences are also very important for more complex stream network characteristics. Also, mainly because of the differences *between* regions, drainage density is a very useful parameter for practical hydrological applications such as regional flood frequency (or low flow) analysis. Here, as in the case of sophisticated stream network theories, the rationale is the complex interplay and interdependence between runoff processes and stream network evolution (Troch *et al.*, 1995). For future research it may be more rewarding to look at the differences *between* different network types as related to geology and climate. Such analyses are more likely to give us a clue on feedback mechanisms and may be more useful for practical applications.

Hydraulic geometry

Hydraulic geometry refers to the quantitative expression of the relationships between the hydraulic characteristics of a channel cross-section and the flow through the channel. Initial work on hydraulic geometry was performed by Leopold and Maddock (1953), who derived a number of empirical relationships based on field measurements. They expressed these relationships in the following way:

$$v \propto Q^m; \quad d \propto Q^f; \quad w \propto Q^b; \quad S \propto Q^z; \quad n \propto Q^y \tag{17}$$

where v is the velocity of flow, d is the mean depth of the flow, w is the surface flow width, S is the energy slope (approximately equal to stream bed slope), n is the Manning friction coefficient and the exponents m, f, b, z and y are constants. These relationships apply to the variations of the hydraulic characteristics with flow, either at a single site, or in the downstream direction (under the assumption that all cross-sections are experiencing flow having an identical probability of exceedance). The exponents m, f, b, z and y would be different for these two instances. Because of the direct connection of discharge to catchment area, these downstream hydraulic geometry relationships can be viewed as scaling relationships. However, caution has to be exercised in such treatment of hydraulic geometry as changes in hydraulic geometry due to floodplain formation and river meandering, often found at large scales, have not been incorporated into these relationships. For example, in large lowland rivers, channels can actually narrow downstream even as the total runoff increases.

Typical values found by Leopold and Maddock (1953) for the exponents for at a site hydraulic geometry were $m = 0{\cdot}34, f = 0{\cdot}40$ and $b = 0{\cdot}26$ ($m + f + b = 1$ since $Q = v \cdot d \cdot w$). For downstream hydraulic geometry, the values were $m = 0{\cdot}1, f = 0{\cdot}4$ and $b = 0{\cdot}5$. Leopold and Langbein (1962) carried out a theoretical analysis of hydraulic geometry based on considerations of entropy and obtained roughly similar values. The small value of m obtained for downstream hydraulic geometry implies that, whereas flow velocity varies considerably with the flow at a single site (i.e. varies in time), it remains approximately constant in the downstream direction (i.e. is almost constant in space). This result has been independently confirmed by a number of workers (Pilgrim, 1977). However, the power law forms of the hydraulic geometry relationships derived by Leopold and Maddock (1953) have been shown to be simplistic in nature, both as regards the at a site and downstream hydraulic geometries they are trying to express. At a site geometry exponents display a functional relationship with discharge expressed by Richards (1982) in log-quadratic form. Furthermore, Ferguson (1973) and Knighton (1974; 1975) show that at a site hydraulic geometry is multivariate rather than bivariate with the sedimentology of the banks playing an important part in the determination of channel width. Downstream hydraulic geometry is also complicated by factors such as spatial variability in the magnitude and frequency of events, especially with regard to upstream and downstream reaches, together with systematic variation at the local scale and the multivariate nature of the controls on channel geometry. Reviews on factors determining the hydraulic geometry relationships can be found in Richards (1982), Knighton (1984) and Hey (1988).

Rodríguez-Iturbe *et al.* (1992b) showed that the combination of three principles of energy expenditure can be used to derive the most important structural characteristics observed in stream networks, including downstream hydraulic geometry variation, the downstream variation of stream slope as a function of catchment area and Horton order ratios. The three principles used by Rodríguez-Iturbe *et al.* (1992b) are: (1) the principle of minimum expenditure in any link of the network; (2) the principle of equal energy expenditure per unit area of channel anywhere in the network; and (3) the principle of minimum total energy expenditure in the network as a whole. However, some of the assumptions, such as a constant rating curve across the catchment, appear to be speculative and need further examination. Chang (1979; 1982), Yang (1976) and Yang and Song (1979) have foreshadowed the work of Rodríguez-Iturbe *et al.* (1992b) by their use of variational principles regarding minimum stream power per unit length and minimum rate of energy dissipation in deriving hydraulic geometries of rivers which have reached stable, equilibrium values, i.e. are in regime.

Geomorphologic instantaneous unit hydrograph

The concept of a geomorphologic instantaneous unit hydrograph (GIUH) was introduced by Rodríguez-Iturbe and Valdés (1979), based on Lee and Delleur (1972; 1976), and was later generalized by Gupta *et al.* (1980) and Rinaldo *et al.* (1991). The original development was geared towards deriving synthetic unit hydrographs for ungauged catchments, but its main use has since been to understand the scaling behaviour of catchment response rather than practical applications. The GIUH is built on the following premises:

1. The GIUH of a catchment can be taken to be the probability distribution of arrival times at the outlet due to a unit impulse of excess rainfall which is uniform across the catchment.
2. The distribution of arrival times is strongly governed by the distribution of water flow pathways from the sources of runoff generation to the catchment outlet.
3. The pathways can be characterized by two fundamental properties. These are (i) a probability that a droplet falling into the catchment will follow this pathway and (ii) a characteristic length representing the distance the droplet will travel from the source of the pathway to the outlet.
4. A pathway is generic — it does not define a path between a specific point on the catchment to the outlet, but rather it defines a set of such points whose individual paths are 'similar', a consequence of the postulates attributed to Shreve (1966; 1967).
5. Each pathway is associated with a pdf of the travel times along that pathway.

Current formulations of the GIUH differ in the manner in which the network morphology is used to characterize the possible pathways and to estimate the probabilities and lengths associated with these pathways. They also differ in the parameterizations assumed to describe the travel time distributions in each pathway. Two alternative approaches have been proposed to characterize the distribution of pathways in terms of network morphology (Snell and Sivapalan, 1994a): (i) methods based on Strahler ordering (Rodríguez-Iturbe and Valdés, 1979; Gupta *et al.*, 1980; Rinaldo *et al.*, 1991); and (ii) methods based on the width and area functions (Lee and Delleur, 1976; Kirkby, 1976; Mesa and Mifflin, 1986; Beven and Wood, 1993; Snell and Sivapalan, 1994a).

In the methods based on Strahler ordering, the concept of a pathway becomes a set of transitions between the initial order of a water droplet (the order of the stream into which the droplet is initially injected) and higher and higher ordered streams until the outlet is eventually reached. Rodríguez-Iturbe and Valdés (1979) and Gupta *et al.* (1980) estimated the pathway probabilities and path lengths in terms of the Horton order ratios. In this formulation the GIUH suppresses the unique structure of an individual network in favour of the expected characteristics of the ensemble of networks that are similar in terms of the Horton order ratios (Surkan, 1969; Beven *et al.*, 1988). Rinaldo *et al.* (1991) showed that, with complete knowledge of the morphology and topology of a Strahler ordered network, these parameters can also be estimated, without recourse to Horton order ratios.

In the methods based on the width or area function, the concept of a pathway as a set of states is not meaningful and consequently we are no longer concerned about the probabilities of transitions from lower to the higher states. Instead, the normalized area function (following division by catchment area), represents the probability of a droplet being sourced at that flow distance from the outlet. It can therefore be considered to be the probability of a droplet finding its way to the outlet through a pathway whose length is that distance.

A number of alternative parameterizations have been suggested to describe travel time distributions for individual flow pathways. Each of these corresponds to an approximate conceptualization of the underlying hydraulics of flow in river channels. These are: (i) dirac delta function, corresponding to pure translation routing (Kirkby, 1976; Gupta *et al.*, 1986b); (ii) exponential distribution, corresponding to a linear store, or series of linear stores (Rodríguez-Iturbe and Valdés, 1979; Gupta *et al.*, 1986b); (iii) the semi-Gaussian Green's function (Mesa and Mifflin, 1986; Troutman and Karlinger, 1985), which corresponds to the linearized diffusion wave equation; (iv) the gamma distribution (Gupta *et al.*, 1986b; van der Tak and Bras, 1990); and (v) the solution to the linearized dynamic wave equation (Harley, 1967; Troutman and Karlinger, 1985).

The GIUH based on Strahler ordering (Gupta *et al.*, 1980; Rinaldo *et al.*, 1991) and that based on the normalized area function (Mesa and Mifflin, 1986; Snell and Sivapalan, 1994a) can be written as

$$\text{Strahler ordering}: \quad f(t) = \sum_{\gamma \in \Gamma} p(\gamma)\cdot h(L_\gamma, t) \tag{18a}$$

$$\text{Area function}: \quad f(t) = \int_0^\infty p(s)\cdot h(s,t)\mathrm{d}s \tag{18b}$$

In Equation (18a), $f(t)$ is the GIUH, γ denotes a particular flow pathway based on Strahler ordering, Γ is the set of all such pathways, $p(\gamma)$ and L_γ are, respectively, the associated pathway probability and pathway length. In Equation (18b), s denotes the flow distance from the outlet and $p(s)$, where $p(s) = a(s)/A$, is the normalized area function. In both equations $h(x,t)$ is an appropriate travel time distribution for a selected flow pathway.

It is clear from the above [Equation (18)] that the two GIUH approaches are indeed comparable, provided the same function $h(x,t)$ is used in both instances. In one instance the result is an integral (area function approach), whereas in the other it is in the form of a finite sum (Strahler ordering). Snell and Sivapalan (1994a) estimated the GIUHs for two natural catchments using the two different approaches and using the semi-Gaussian Green's function of Mesa and Mifflin (1986) to parameterise $h(x,t)$. They found that the approach based on Strahler ordering produced a GIUH which was substantially different from that produced by the area function approach. This is because Strahler ordering narrows down the number of possible pathways (to five or 10, for example) which narrows down the possibilities, while the area function allows an infinite number of pathways.

As the GIUH is either parameterized in terms of the Horton order ratios (which are dimensionless quantities) or the normalized area function (which is also dimensionless), the GIUH can be viewed as a scaling relationship and can be used to link the responses of catchments of different sizes. For example, Rodríguez-Iturbe and Valdés (1979) showed that the non-dimensional peak, q_p^*, and time to peak, t_p^*, of the GIUH can be approximated by

$$q_p^* = \frac{q_p L_\Omega}{v} = 1{\cdot}31 R_L^{0{\cdot}43} \tag{19a}$$

$$t_p^* = \frac{t_p v}{L_\Omega} = 0{\cdot}44 R_L^{-0{\cdot}38} (R_B/R_A)^{0{\cdot}55} \tag{19a}$$

where L_Ω is the length of the highest order stream in the network (the characteristic length scale) and v is a velocity parameter. Thus two catchments would be similar, in terms of their GIUH and regardless of their size, if R_L and R_B/R_A are the same. Equations (19a) and (19b) will then serve as the scaling relationships.

The GIUH approach described so far has ignored the processes of runoff generation on the hillslopes, implicitly assuming that runoff is generated uniformly over the entire catchment area and that hillslope travel times are negligible compared with the travel times within the channel network. Sivapalan *et al.* (1990) have generalized the GIUH of Rodríguez-Iturbe and Valdés (1979) to incorporate partial area runoff generation. Lee and Delleur (1976), Kirkby (1976) and Mesa and Mifflin (1986) have presented a general framework which can be used to extend the area function approach so as to incorporate hillslope response functions.

Another limitation is that the GIUH is a linear construct, underlain by the assumptions of constant velocity and independence of flow pathways. Although this may be an adequate approximation for large catchments, empirical evidence suggests that small catchments arer highly non-linear, with the non-linearity decreasing with increasing catchment size (Minshall, 1960). This may be a consequence of the fact that channel processes are not as important in small catchments. The GIUH, being a linear model, may not therefore be applicable to small catchments. The quantification of the non-linearity with decreasing catchment size, and the investigation of the sources of non-linearity (hillslopes versus channel network) are important problems that cannot be addressed by the GIUH approach.

Meta-channel concept

As a way of overcoming such limitations of the GIUH approach, Snell and Sivapalan (1995) made a first step with the introduction of their *meta-channel* concept. They form their meta-channel by collapsing the stream network on the basis of flow distance and estimate the hydraulic properties of the new meta-channel by conserving the spatial distributions (in terms of flow distance) of mass, momentum and energy between the natural network and the meta-channel.

Their development of the meta-channel precludes the *a priori* need to assume linearity of channel behaviour. Rather, physically based governing equations, such as the St. Venant equations, can be used to model the flow. Computationally, the modelling of the hydraulics is much simpler as there is only a single channel to model, whose hydraulic characteristics are already known. It is also fairly straightforward to integrate more realistic hillslope response functions with the meta-channel approach. The disadvantage is that the meta-channel approach cannot yield an analytical solution. However, it does give a more general, non-linear solution which can be used as a benchmark against which the accuracy of more approximate solutions can be tested. More specifically, the metal-channel approach permits the development of scaling relationships that hold for a wider range of catchment sizes, as compared with the GIUH.

FUTURE RESEARCH

This paper has defined a framework for scale issues in hydrology. As a starting point, some basic definitions and fundamental issues of heterogeneity were discussed. Then two alternative approaches to scaling were suggested. The first is a model-oriented approach and focuses on the scaling of state variables, model parameters, inputs and conceptualizations. The second (i.e. dimensional analysis and similarity concepts) deals with complex processes in a much simpler fashion. In each of these areas considerable advances have been achieved in the last years. There are, however, a number of key issues that need more attention in the future:

1. Organization in catchments has been identified as one of the keys to 'enlightened scaling'. New and innovative techniques for quantifying organization are needed.
2. The space resolution of hydrological measurements is typically much poorer than the time resolution. What is needed is high-resolution spatial data of quantities such as soil moisture. Also, the data should be representative over the soil profile (or at least over a depth of several decimetres), rather than just over a thin layer at the soil surface. These patterns are needed for defining scale relationships, identifying organization and evaluating distributed parameter models.
3. Current work on stream network analysis has often neglected geologic information. Geology is clearly a dominant control of network structure and should therefore play a central part in the analysis. It may be more rewarding to search for relationships between geology and network characteristics than to refine theories that are sophisticated but disregard the most obvious controls.
4. Theoretical achievements in hydrological scaling have been considerable over the last few years and decades, but, often, they have not contributed much to solve the problems of engineering hydrology. There is a definite need for more sophisticated techniques in practice and we trust that adapting current theoretical concepts to the engineering practice is a feasible endeavour. Indeed, for future scale research in hydrology it will be one of the most rewarding challenges to bridge the gap between theory and practice.

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REFERENCES

Ababou, R. and Wood, E. F. 1990. 'Comment on "Effective groundwater model parameter values: Influence of spatial variability of hydraulic conductivity, leakance, and recharge" by J. J. Gómez-Hernández and S. M. Gorelick', *Wat. Resour. Res.*, **26**, 1843–1846.

Abbott, M. B., Bathurst, J. C., Cunge, J. A., O'Connell, P. E., and Rasmussen, J. 1986. 'An introduction to the European Hydrological System — Système Hydrologique Européen, "SHE", 2: structure of a physically-based, distributed modelling system', *J. Hydrol.*, **87**, 61–77.

Abrahams, A. D., Parsons, A. J., and Luk, S. H. 1989. 'Distribution of depth of overland flow on desert hillslopes and its implications for modelling soil erosion', *J. Hydrol.*, **106**, 177–184.

Allen, T. F. H., and Starr, T. B. 1982. *Hierarchy*. The University of Chicago Press, Chicago. 310 pp.

Anderson, M. G. (Ed.) 1988. *Modelling Geomorphological Systems*. Wiley, Chichester. 458 pp.

Anderson, M. G. and Burt, T. P. 1990. 'Subsurface runoff' in Anderson, M. G. and Burt, T. P. (Eds), *Process Studies in Hillslope Hydrology*. Wiley, Chichester. pp. 365–400.

Anderson, M. P. 1989. 'Hydrogeologic facies models to delineate large-scale spatial trends in glacial and glaciofluvial sediments', *Geol. Soc. Am. Bull.*, **101**, 501–511.

Anderson, M. P. 1991. 'Comment on "Universal scaling of hydraulic conductivities and dispersivities in geologic media" by S. P. Neuman', *Wat. Resour. Res.*, **27**, 1381–1382.

Austin, P. M. and Houze, R. A. 1972. 'Analysis of the structure of precipitation patterns in New England', *J. Appl. Meteorol.*, **11**, 926–935.

Avissar, R. 1995. 'Scaling of land–atmosphere interactions: an atmospheric modelling perspective', *Hydrol. Process.*, **9**, 000–000.

Bakr, A. A., Gelhar, L. W., Gutjahr, A. L., and MacMillan, J. R. 1978. 'Stochastic analysis of spatial variability in subsurface flows. 1. Comparison of one- and three-dimensional flows', *Wat. Resour. Res.*, **14**, 263–271.

Band, L. E. 1986. 'Topographic partition of watersheds with digital elevation models', *Wat. Resour. Res.*, **22**, 15–24.

Band, L. E. and Moore, I. D. 1995. 'Scale: landscape attributes and geographical information systems', *Hydrol. Process.*, **9**, 000–000.

Barancourt, C., Creutin, J. D., and Rivoirard, J. 1992. 'A method for delineating and estimating rainfall fields', *Wat. Resour. Res.*, **28**, 1133–1144.

Barling, R. D., Moore, I. D., and Grayson, R. B. 1994. 'A quasi-dynamic wetness index for characterising the spatial distribution of zones of surface saturation and soil water content', *Wat. Resour. Res.*, **30**, 1029–1044.

Bathurst, J. C. and Cooley, K. R. 'Use of the SHE hydrological modelling system to investigate basin response to snowmelt at Reynolds Creek, Idaho', *J. Hydrol.*, in press.

Benecke, P. 1992. 'Vorhersagbarkeit der Wasserbindungs- und der Wasserleitfähigkeitsfunktion an bodenkundlichen Substratmerkmalen' in Kleeberg, H.-B. (Ed.), *Regionalisierung in der Hydrologie, DFG-Mitt. XI.* VCH Verl. ges., Weinheim. pp. 221–239.
Bergström, S. 1991. 'Principles and confidence in hydrological modelling', *Nordic Hydrol.*, **22,** 123–136.
Beven, K. 1981. 'Kinematic subsurface stormflow', *Wat. Resour. Res.*, **17,** 1419–1424.
Beven, K. 1986. 'Runoff production and flood frequency in catchments of order n: an alternative approach' in Gupta, V. K., Rodríguez-Iturbe, I., and Wood, E. F. (Eds), *Scale Problems in Hydrology.* D. Reidel, Dordrecht. pp. 107–131.
Beven, K. 1989. 'Changing ideas in hydrology — the case of physically based models', *J. Hydrol.*, **105,** 157–172.
Beven, K. J. 1991. 'Scale considerations' in Bowles, D. S. and O'Connell, P. E. (Eds), *Recent Advances in the Modeling of Hydrologic Systems.* Kluwer, Dordrecht. pp. 357–371.
Beven, K. (Ed.) 1992. 'Future of distributed modelling', *Hydrol. Proc.*, **6,** 253–268.
Beven, K. J. and Kirkby, M. J. 1979. 'A physically-based variable contributing area model of basin hydrology', *Hydrol. Sci. Bull.*, **24,** 43–69.
Beven, K. and Wood, E. F. 1993. 'Flow routing and the hydrological response of channel networks' in Beven, K. and Kirkby, M. J. (Eds), *Channel Network Hydrology.* Wiley, Chichester. pp. 99–128.
Beven, K., Warren, R., and Zaoui, J. 1980. 'SHE: towards a methodology for physically-based distributed forecasting in hydrology', *Proc. Oxford Symp., IAHS Publ.*, **129,** 133–137.
Beven, K., Wood, E. F., and Sivapalan, M. 1988. 'On hydrological heterogeneity — catchment morphology and catchment response', *J. Hydrol.*, **100,** 353–375.
Binley, A., Beven, K., and Elgy, J. 1989. 'A physically based model of heterogeneous hillslopes. 2. Effective hydraulic conductivities', *Wat. Resour. Res.*, **25,** 1227–1233.
Blöschl, G., Kirnbauer, R., and Gutknecht, D. 1991. 'Distributed snowmelt simulations in an Alpine catchment. 1. Model evaluation on the basis of snow cover patterns', *Wat. Resour. Res.*, **27,** 3171–3179.
Blöschl, G., Gutknecht, D., Grayson, R. B., Sivapalan, M., and Moore, I. D. 1993. 'Organisation and randomness in catchments and the verification of hydrologic models', *EOS, Trans. Am. Geophys. Union*, **74,** 317.
Blöschl, G., Gutknecht, D., and Kirnbauer, R. 1994. 'On the evaluation of distributed hydrologic models' in Rosso, R., Peano, A., Becchi, I., Bemporad, G. A. (Eds.), *Advances in Distributed Hydrology. Proceedings of a Workshop held in Bergamo, Italy, June 1992.* Water Resources Publications, in press.
Blöschl, G., Grayson, R. B., and Sivapalan, M. 1995. 'On the representative elementary area (REA) concept and its utility for distributed rainfall-runoff modelling', *Hydrol. Process.*, **9,** 000–000.
Blöschl, G., Gutknecht, D., and Grayson, R. B. 'On spatial organisation and randomness in hydrology', *Wat. Resour. Res.*, submitted.
Brannan, J. R. and Haselow, J. S. 1993. 'Compound random field models of multiple scale hydraulic conductivity', *Wat. Resour. Res.*, **29,** 365–372.
Bridges, E. M. 1982. 'Techniques of modern soil survey' in Bridges, E. M. and Davidson, D. A. (Eds), *Principles and Applications of Soil Geography.* Longman, London, New York. pp. 29–57.
Buckingham, E. 1914. 'On physically similar systems: Illustrations of the use of dimensional equations', *Phys. Rev.*, **4,** 345–376.
Burt, T. P. and Butcher, D. P. 1985. 'Topographic controls of soil moisture distributions', *J. Soil Sci.*, **36,** 469–486.
Canterford, R. P., Pescod, N. R., Pearce, H. J., and Turner, L. H. 1987. 'Design intensity–frequency–duration rainfall' in Pilgrim, D. H. (Ed.), *Australian Rainfall and Runoff.* The Institution of Engineers, Barton, ACT. pp. 15–40.
Carlson, A. B. 1986. *Communication Systems.* 3rd edn. McGraw Hill, New York. 686pp.
Chang, H. H. 1979. 'Minimum stream power and river channel patterns', *J. Hydrol.*, **41,** 303–327.
Chang, H. H. 1982. 'Mathematical model for erodible channels', *J. Hydraul. Div. ASCE*, **108,** (HY5), 678–689.
Chappell, N. and Ternan, L. 1992. 'Flow path dimensionality and hydrological modelling', *Hydrol. Process.*, **6,** 327–345.
Chhabra, A. B., Jensen, R. V., and Sreenivasan, K. R. 1989. 'Extraction of underlying multiplicative processes from multifractals via the thermodynamic formalism', *Phys. Rev. A*, **40,** 4593–4611.
Chorley, R. J., Schumm, S. A., and Sugden, D. E. 1984. *Geomorphology.* Methuen, London. 605 pp.
Chow, V. T., Maidment, D. R., and Mays, L. W. 1988. *Applied Hydrology.* McGraw-Hill, New York. 572 pp.
Clark, W. 1985. 'Scales of climate impacts', *Climatic Change*, **7,** 5–27.
Copty, N., Rubin, Y., and Mavko, G. 1993. 'Geophysical–hydrological identification of field permeabilities through Bayesian updating', *Wat. Resour. Res.*, **29,** 2813–2825.
Creutin, J. D. and Obled, C. 1982. 'Objective analyses and mapping techniques for rainfall fields: an objective comparison', *Wat. Resour. Res.*, **18,** 413–431.
Cushman, J. H. 1983. 'Comment on "Three-dimensional stochastic analysis of macrodispersion in aquifers" by L. W. Gelhar and C. L. Axness', *Wat. Resour. Res.*, **19,** 1641–1642.
Cushman, J. H. 1984. 'Unifying the concepts of scale, instrumentation, and stochastics in the development of multiphase transport theory', *Wat. Resour. Res.*, **20,** 1668–1676.
Cushman, J. H. 1987. 'More on stochastic models', *Wat. Resour. Res.*, **23,** 750–752.
Cutler, C. D. 1993. 'A review of the theory and estimation of fractal dimension', *Tech. Rep. Ser. STAT-93-06*, Department of Statistics and Actuarial Science, University of Waterloo, Ontario, 107 pp.
Dagan, G. 1979. 'Models of groundwater flow in statistically homogeneous porous formations', *Wat. Resour. Res.*, **15,** 47–63.
Dagan, G. 1986. 'Statistical theory of groundwater flow and transport: pore to laboratory, laboratory to formation and formation to regional scale', *Wat. Resour. Res.*, **22,** 120S–134S.
Daubechies, I. 1992. *Ten Lectures on Wavelets.* SIAM, Philadelphia. 357 pp.
de Boer, D. H. 1992. 'Hierarchies and spatial scale in process geomorphology: a review', *Geomorphology*, **4,** 303–318.
de Marsily, G. 1986. *Quantitative Hydrogeology.* Academic Press, San Diego. 440 pp.
Denbigh, K. G. 1975. 'A non-conserved function for organized systems' in Kubát, L. and Zeman, J. (Eds), *Entropy and Information in Science and Philosophy.* Elsevier, Amsterdam. pp. 83–92.

Deutsch, A. and Journel, A. G. 1992. *GSLIB, Geostatistical Software Library and User's Guide*. Oxford University Press, New York, Oxford. 340 pp.
deVriend, H. J. 1991. 'Mathematical modelling and large-scale coastal behaviour', *J. Hydraul. Res.*, **29**, 727–740.
Dietrich, W. E. and Dunne, T. 1993. 'The channel head' in Beven, K. and Kirkby, M. J. (Eds), *Channel Network Hydrology*. Wiley, Chichester. pp. 175–219.
Dingman, S. L. 1994. *Physical Hydrology*. Macmillan, New York. 575 pp.
Dooge, J. C. I. 1982. 'Parameterization of hydrologic processes' in Eagleson, P. S. (Ed.), *Land Surface Processes in Atmospheric General Circulation Models*. Cambridge University Press, London. pp. 243–288.
Dooge, J. C. I. 1986. 'Looking for hydrologic laws', *Wat. Resour. Res.*, **22**, 46S–58S.
Dovers, S. R. 1995. 'A framework for scaling and framing policy problems in sustainability', *Ecol. Econ.*, **12**, 93–106.
Dozier, J. 1992. 'Opportunities to improve hydrologic data', *Rev. Geophys.*, **30**, 315–331.
Drury, S. A. 1987. *Image Interpretation in Geology*. Allen and Unwin, London. 243 pp.
Dunne, T. 1978. 'Field studies of hillslope flow processes' in Kirkby, M. J. (Ed.), *Hillslope Hydrology*. Wiley, Chichester. pp. 227–293.
Dunne, T. 1983. 'Relation of field studies and modeling in the prediction of storm runoff', *J. Hydrol.*, **65**, 25–48.
Dykaar, B. B. and Kitanidis, P. K. 1992. 'Determination of the effective hydraulic conductivity for heterogeneous porous media using a numerical spectral approach. 2. Results', *Wat. Resour. Res.*, **28**, 1167–1178.
Eagleson, P. S. 1972. 'Dynamics of flood frequency', *Wat. Resour. Res.*, **8**, 878–898.
Eagleson, P. S. 1978. 'Climate, soil, and vegetation (in 7 parts)', *Wat. Resour. Res.*, **14**, 705–776.
El-Kadi, A. and Brutsaert, W. 1985. 'Applicability of effective parameters for unsteady flow in nonuniform aquifers', *Wat. Resour. Res.*, **21**, 183–198.
England, C. B. and Holtan, H. N. 1969. 'Geomorphic grouping of soils in watershed engineering', *J. Hydrol.*, **7**, 217–225.
Engman, E. T. 1986. 'Roughness coefficients for routing surface runoff', *J. Irrig. Drainage Div. Proc. ASCE*, **112**, 39–53.
Engman, E. T. and Gurney, R. J. 1991. 'Recent advances and future implications of remote sensing for hydrologic modeling' in Bowles, D. S. and O'Connell, P. E. (Eds), *Recent Advances in the Modeling of Hydrologic Systems*. Kluwer, Dordrecht. pp. 471–495.
Falconer, K. J. 1990. *Fractal Geometry: Mathematical Foundations and Applications*. Wiley, Chichester. 288 pp.
Famiglietti, J. S. 1992. 'Aggregation and scaling of spatially-variable hydrological processes: local, catchment-scale and macroscale models of water and energy balance, *PhD Dissertation*, Princeton University, 207 pp.
Feder, J. 1988. *Fractals*. Plenum Press, New York and London. 283 pp.
Ferguson, R. I. 1973. 'Channel pattern and sediment type', *Area*, **5**, 38–41.
Fischer, H. B., List, E. J., Koh, R. C. Y., Imberger, J., and Brooks, N. H. 1979. *Mixing in Inland and Coastal Waters*. Academic Press, New York, 483 pp.
Fitzharris, B. B. 1975. 'Snow accumulation and deposition on a westcoast midlatitude mountain', *PhD Thesis*, University of British Columbia, Vancouver 367 pp.
Flood Studies Report 1975. *Vol. I — Hydrological Studies*. Natural Environment Research Council, London. 570 pp.
Fortak, H. 1982. *Meteorologie*. Dietrich Reimer, Berlin. 293 pp.
Foster, G. R., Huggins, L. F., and Meyer, L. D. 1984. 'A laboratory study of rill hydraulics: I. Velocity relationships', *Trans. Am. Soc. Agric. Engin.*, **27**, 790–796.
Foufoula-Georgiou, E. and Georgakakos, K. P. 1991. 'Hydrologic advances in space–time precipitation modeling and forecasting' in Bowles, D. S. and O'Connell, P. E. (Eds), *Recent Advances in the Modeling of Hydrologic Systems*. Kluwer, Dordrecht. pp. 47–65.
Freeze, R. A. 1975. 'A stochastic-conceptual analysis of one-dimensional groundwater flow in nonuniform homogeneous media', *Wat. Resour. Res.*, **11**, 725–741.
Freeze, R. A. and Harlan, R. L. 1969. 'Blueprint for a physically-based, digitally simulated hydrologic response model', *J. Hydrol.*, **9**, 237–258.
Gallant, J. C., Moore, I. D., Hutchinson, M. F., and Gessler, P. E. 1994. 'Estimating fractal dimension of profiles: a comparison of methods', *Math. Geol.*, **26**, 455–481.
Gelhar, L. W. 1986. 'Stochastic subsurface hydrology from theory to applications', *Wat. Resour. Res.*, **22**, 135S–145S.
Germann, P. F. 1986. 'Rapid drainage response to precipitation', *Hydrol. Process.*, **1**, 3–14.
Germann, P. F. 1990. 'Macropores and hydrologic hillslope processes' in Anderson, M. G. and Burt, T. P. (Eds), *Process Studies in Hillslope Hydrology*. Wiley, Chichester. pp. 327–367.
Gessler, P. E., McKenzie, N. J., Hutchinson, M. F., and Moore, I. D. 1993. 'Soil–landscape modelling in southeastern Australia: scale relationships' in Kalma, J., Sivapalan, M., and Wood, E. (Eds), *Scale Issues in Hydrological/Environmental Modelling, Proc. Workshop Robertson, CRES*. Australian National University, Canberra. p. 9.
Golding, D. L. 1974. 'The correlation of snowpack with topography and snowmelt runoff on Marmot Creek basin, Alberta', *Atmosphere*, **12**, 31–38.
Gómez-Hernández, J. J. and Gorelick, S. M. 1989. 'Effective groundwater model parameter values: influence of spatial variability of hydraulic conductivity, leakance, and recharge', *Wat. Resour. Res.*, **25**, 405–419.
Goodrich, D. C. 1990. 'Geometric simplification of a distributed rainfall–runoff model over a large range of basin scales', *PhD Thesis*, Univ. Arizona, Tucson, 361 pp.
Goodrich, D. C. and Woolhiser, D. A. 1991. 'Catchment hydrology', *Rev. Geophys.* (suppl.), April, 202–209.
Grace, R. A., and Eagleson, P. S. 1966. 'The synthesis of short-time-increment rainfall sequences', *Rep. No. 91*, MIT, Hydrodynamics Laboratory 105 pp.
Grayson, R. B., Moore, I. D., and McMahon, T. A. 1992. 'Physically-based hydrologic modelling: 2. Is the concept realistic?', *Wat. Resour. Res.*, **26**, 2659–2666.
Grayson, R. B., Blöschl, G., Barling, R. D., and Moore, I. D. 1993. 'Process, scale and constraints to hydrological modelling in GIS' in Kovar, K. and Nachtnebel, H. P. (Eds), *Applications of Geographic Information Systems in Hydrology and Water Resources Management (Proc. Vienna Symp., April 1993), IAHS Publ.*, **211**, 83–92.

Grayson, R. B., Blöschl, G., and Moore, I. D. 'Distributed parameter hydrologic modelling using vector elevation data: THALES and TAPES-C' in Singh, V. P. (Ed.), *Computer Models of Watershed Hydrology*. CRC Press, in press.
Green, W. H. and Ampt, G. A. 1911. 'Studies on soil physics, Part I. The flow of air and water through soils', *J. Agric. Sci.*, **4**, 1–24.
Guerra, L., Moore, I. D., Kalma, J. D., and Hofstee, C. 1993. 'Predicting spatially distributed evaporation using terrain, soil and land cover information' in Bolle, H.-J., Feddes, R. A., and Kalma, J. D. (Eds), *Exchange Processes at the Land Surface for a Range of Space and Time Scales, IAHS Publ.*, **212**, 611–618.
Gumbel, E. J. 1941. 'The return period of flood flows', *Ann. Math. Stat.*, **12**, 163–190.
Gupta, V. K. and Dawdy, D. R. 'Regional analysis of flood peaks: multiscaling theory and its physical basis', in Rosso, R., Peano, A., Becchi, I., Bemporad, G. A. (Eds.), *Advances in Distributed Hydrology. Proceedings of a Workshop held in Bergamo, Italy, June 1992.* Water Resources Publications, pp. 149–168.
Gupta, V. K. and Dawdy, D. R. 1995. 'Some physical implications of regional variations in the scaling exponents of flood quantiles', *Hydrol. Process.*, **9**, 347–361.
Gupta, V. K. and Waymire, E. C. 1990. 'Multiscaling properties of spatial rainfall and river flow distributions', *J. Geophys. Res.*, **95**(D3), 1999–2009.
Gupta, V. K. and Waymire, E. C. 1993. 'A statistical analysis of mesoscale rainfall as a random cascade', *J. Appl. Meteorol.*, **32**, 251–267.
Gupta, V. K., Waymire, E., and Wang, J. R. 1980. 'A representation of an instantaneous unit hydrograph from geomorphology', *Wat. Resour. Res.*, **16**, 855–862.
Gupta, V. K., Rodríguez-Iturbe, I., and Wood, E. F. (Eds) 1986a. *Scale Problems in Hydrology*. D. Reidel, Dordrecht. 246 pp.
Gupta, V. K., Waymire, E., and Rodríguez-Iturbe, I. 1986b. 'On scales, gravity and network structure' in Gupta, V. K., Rodríguez-Iturbe, I., and Wood, E. F. (Eds), *Scale Problems in Hydrology*. D. Reidel, Dordrecht, pp. 159–184.
Gutjahr, A. L., Gelhar, L. W., Bakr, A. A., and MacMillan, J. R. 1978. 'Stochastic analysis of spatial variability in subsurface flows, 2, evaluation and application', *Wat. Resour. Res.*, **14**, 953–959.
Gutknecht, D. 1991a. 'On the development of "applicable" models for flood forecasting' in Van de Ven, F. H. M., Gutknecht, D., Loucks, D. P., and Salewicz, K. A. (Eds), *Hydrology for the Water Management of Large River Basins (Proc. Vienna Symp., August 1991), IAHS Publ.*, **201**, 337–345.
Gutknecht, D. 1991b. 'Computer aided modelling for operational forecasting systems', *Ann. Geophys. Suppl.* **9**, C480.
Gutknecht, D. 1993. 'Grundphänomene hydrologischer Prozesse', *Zürcher Geogr. Schriften*, **53**, Geographisches Institut der Eidgenossischen Technischen Hochschule, Zurich, 25–38.
Gyasi-Agyei, Y., Willgoose, G., and De Troch, F. P. 1995. 'Effects of vertical resolution and map scale of digital elevation models on geomorphologic parameters used in hydrology', *Hydrol. Process.*, **9**, 363–382.
Hack, J. T. and Goodlett, J. G. 1960. 'Geomorphology and forest ecology of a mountain region in the Central Appalachians', *US Geol. Surv. Prof. Pap.*, **347**, 66pp.
Halsey, T. C., Jensen, M. H., Kadanoff, L. P., Procaccia, I., and Shraiman, B. I. 1986. 'Fractal measures and their singularities: the characterization of strange sets', *Phys. Rev. A*, **33**, 1141–1151.
Haltiner, G. J. and Williams, R. T. 1980. *Numerical Prediction and Dynamic Meteorology*. Wiley, New York. 477 pp.
Harley, B. M. 1967. 'Linear routing in uniform open channels', *M. Eng. Sci. Thesis*, Department of Civil Engineering, National University of Ireland.
Hatton, T. J. and Wu, H. I. 1995. 'Scaling theory to extrapolate individual tree water use to stand water use', *Hydrol. Process.*, **9**, 000–000.
Haury, L. R., McGowan, J. A., and Wiebe, P. H. 1977. 'Patterns and processes in the time–space scales of plankton distributions' in Steele, J. H. (Ed.), *Spatial Pattern in Plankton Communities*. Plenum Press, New York and London. pp. 277–327.
Hebson, C. S. and Wood, E. F. 1986. 'A study of the scale effects in flood frequency response' in Gupta, V. K., Rodríguez-Iturbe, I., and Wood, E. F. (Eds), *Scale Problems in Hydrology*. D. Reidel, Dordrecht. pp. 133–158.
Hellums, J. D. and Churchill, S. W. 1961. 'Dimensional analysis and natural circulation', *Chem. Eng. Progr. Symp. Ser.*, **57**, 75–80.
Helmlinger, K. R., Kumar, P., and Foufoula-Georgiou, E. 1993. 'On the use of digital elevation model data for Hortonian and fractal analyses of channel networks', *Wat. Resour. Res.*, **29**, 2599–2613.
Hey, R. D. 1988. 'Mathematical models of channel morphology' in Anderson, M. G. (Ed.), *Modelling Geomorphological Systems*. Wiley, Chichester. pp. 99–125.
Hillel, D., and Elrick, D. E. 1990. *Scaling in Soil Physics: Principles and Applications*. Soil Science Society of America, Vol. 25. 122 pp.
Hoeksema, R. J., and Kitanidis, P. K. 1985. 'Analysis of the spatial structure of properties of selected aquifers', *Wat. Resour. Res.*, **21**, 563–572.
Hoosbeek, M. R. and Bryant, R. B. 1992. 'Towards the quantitative modeling of pedogenesis — a review', *Geoderma*, **55**, 183–210.
Horton, R. E. 1945. 'Erosional development of streams and their drainage basins: hydrophysical approach to quantitative morphology', *Geol. Soc. Am. Bull.*, **38**, 275–370.
Huff, F. A. 1967. 'Time distribution of rainfall', *Wat. Resour. Res.*, **3**, 1007–1018.
Hurst, H. E. 1951. 'Long-term storage capacity of reservoirs', *Trans. Am. Soc. Civ. Engin.*, **116**, 770–808.
Hutchinson, M. F. 1991. 'The application of thin plate smoothing splines to continent-wide data assimilation' in Jasper, J. D. (Ed.), *Data Assimilation Systems, BMRC Res. Rep. 27*, Bureau of Meteorology, Melbourne: 104–113.
Ijjasz-Vasquez, E. J., Rodríguez-Iturbe, I., and Bras, R. L. 1992. 'On the multifractal characterization of river basins', *Geomorphology*, **5**, 297–310.
Jenkins, G. M. and Watts, D. G. 1968. *Spectral Analysis and its Applications*. Holden-Day, San Francisco. 525 pp.
Jenny, H. 1941. *Factors of Soil Formation*. McGraw Hill, New York. 281 pp.
Jenny, H. 1980. *The Soil Resource*. Springer, New York. 377 pp.
Jensen, H. 1989. 'Räumliche Interpolation der Stundenwerte von Niederschlag, Temperatur und Schneehöhe', *Zürcher. Geogr. Schriften*, **35**, Geographisches Institut der Eidgenössischen Technischen Hochschule, Zurich, 70 pp.
Jenson, S. K. and Domingue, J. O. 1988. 'Extracting topographic structure from digital elevation data for geographic information system analysis', *Photogr. Engin. Remote Sensing*, **54**, 1593–1600.

Jones, J. A. A. 1987. 'The effect of soil piping on contributing areas and erosion patterns', *Earth Surf. Process. Landforms*, **12**, 229–248.
Journel, A. G. 1986. 'Constrained interpolation and qualitative information — the soft kriging approach', *Math. Geol.*, **18**, 269–287.
Journel, A. G. and Huijbregts, C. J. 1978. *Mining Geostatistics*. Academic Press, London. 600 pp.
Kern, K. 1994. *Grundlagen naturnaher Gewässergestaltung*. Springer, Berlin. 256 pp.
Kirchner, J. W. 1993. 'Statistical inevitability of Horton's laws and the apparent randomness of stream channel networks', *Geology*, **21**, 591–594.
Kirkby, M. J. 1976. 'Tests of the random network model and its applications to basin hydrology', *Earth Surf. Process.*, **1**, 197–212.
Kirnbauer, R., Blöschl, G., and Gutknecht, D. 1994. 'Entering the era of distributed snow models', *Nordic Hydrol.*, **25**, 1–24.
Kleeberg, H.-B. (Ed.) 1992. *Regionalisierung in der Hydrologie. DFG-Mitt. XI.* VCH Verl. ges., Weinheim. 444 pp.
Klemeš, V. 1983. 'Conceptualisation and scale in hydrology', *J. Hydrol.*, **65**, 1–23.
Klinkenberg, B. and Goodchild, M. F. 1992. 'The fractal properties of topography: a comparison of methods', *Earth Surf. Process. Landforms*, **17**, 217–234.
Kneale, W. R., and White, R. E. 1984. 'The movement of water through cores of a dry (cracked) clay-loam grassland topsoil', *J. Hydrol.*, **67**, 361–365.
Knighton, A. D. 1974. 'Variation in width–discharge relation and some implications for hydraulic geometry', *Bull. Geol. Soc. Am.*, **85**, 1059–1076.
Knighton, A. D. 1975. 'Variations in at-a-station hydraulic geometry', *Am. J. Sci.*, **275**, 186–218.
Knighton, D. 1984. *Fluvial Forms and Processes*. Edward Arnold, London. 218 pp.
Koide, S. and Wheater, H. S. 1992. 'Subsurface flow simulation of a small plot at Loch Chon, Scotland', *Hydrol. Process.*, **6**, 299–326.
Kölla, E. 1987. 'Estimating flood peaks from small rural catchments in Switzerland', *J. Hydrol.*, **95**, 203–225.
Koutsoyiannis, D. and Foufoula-Georgiou, E. 1993. 'A scaling model of a storm hyetograph', *Wat. Resour. Res.*, **29**, 2345–2361.
Krasovskaia, I. 1982. 'Hypothesis on runoff formation in small watersheds in Sweden', *FoU-notiser No. 19*, SMHI, Norrköping.
Kupfersberger, H. 1994. 'Integrating different types of information for estimating aquifer transmissivities with the sequential indicator cosimulation method' in Dracos, T. H. and Stauffer, F. (Eds), *Transport and Reactive Processes in Aquifers (Proc. 5 IAHR Symp. Zurich)*, Balkema, Rotterdam pp. 165–170.
Kupfersberger, H. and Blöschl, G. 1995. 'Estimating aquifer transmissivities — on the value of auxiliary data', *J. Hydrol.*, **165**, 85–99.
Ladson, A. R., and Moore, I. D. 1992. 'Soil water prediction on the Konza Prairie by microwave remote sensing and topographic attributes', *J. Hydrol.*, **138**, 385–407.
Lang, H. 1985. 'Höhenabhängigkeit der Niederschläge' in *Der Niederschlag in der Schweiz, Beiträge zur Geologie der Schweiz, Nr. 31.* Verl. Kümmerlyt. Frey, Bern. pp. 149–157.
Langhaar, H. L. 1951. *Dimensional Analysis and Theory of Models*. Wiley, New York. 166 pp.
Larsen, J. E., Sivapalan, M., Coles, N. A., and Linnet, P. E. 1994. 'Similarity analysis of runoff generation processes in real-world catchments', *Wat. Resour. Res.*, **30**, 1641–1652.
Leavesley, G. H. and Hay, L. E. 1993. 'A nested-model approach for investigating snow accumulation and melt processes in mountainous regions', *EOS, Trans. Am. Geophys. Union*, **74**, 237.
Leavesley, G. H. and Stannard, L. G. 1990. 'Application of remotely sensed data in a distributed-parameter watershed model' in Kite, G. W. and Wankiewicz, A. (Eds), *Proc. Workshop on Applications of Remote Sensing in Hydrology, Saskatoon, February 1990.* pp. 47–64.
Lebel, T., Bastin, G., Obled, C., and Creutin, J. D. 1987. 'On the accuracy of areal rainfall estimation: a case study', *Wat. Resour. Res.*, **23**, 2123–2134.
Lee, M. T. and Delleur, J. W. 1972. 'A program for estimating runoff from Indiana watersheds, 3, analysis of geomorphologic data and a dynamic contributing area model for runoff estimation', *Tech. Rep. 24*, Purdue Univ. Water Resour. Res. Center, Lafayette 144 pp.
Lee, M. T. and Delleur, J. W. 1976. 'A variable source area model of the rainfall–runoff process based on the watershed stream network', *Wat. Resour. Res.*, **12**, 1029–1036.
Leopold, L. B. and Langbein, W. B. 1962. 'The concept of entropy in landscape evolution', *US Geol. Surv. Prof. Pap.*, **500-A**, A20 pp.
Leopold, L. B. and Maddock, T. 1953. 'The hydraulic geometry of stream channels and some physiographic implications', *US Geol. Surv. Prof. Pap.*, **252**, 57 pp.
Lettenmaier, D. P. and Burges, S. J. 1977. 'Operational assessment of hydrologic models of longterm persistence', *Wat. Resour. Res.*, **13**, 113–124.
Lewis, A. 1995. 'Scale in spatial environmental databases', *PhD Thesis*, Centre for Resource and Environmental Studies, Australian National University, Canberra.
Linsley, R. K., Kohler, M. A., and Paulhus, J. L. H. 1988. *Hydrology for Engineers*. McGraw Hill, New York. 492 pp.
Loague, K. 1990. 'R-5 Revisited. 2. Reevaluation of a quasi-physically based rainfall-runoff model with supplemental information', *Wat. Resour. Res.*, **26**, 973–987.
Lovejoy, S. 1982. 'The area–perimeter relations for rain and cloud areas', *Science*, **216**, 185–187.
Lovejoy, S. and Schertzer, D. 1985. 'Generalized scale invariance in the atmosphere and fractal models of rain', *Wat. Resour. Res.*, **21**, 1233–1250.
Lovejoy, S. and Schertzer, D. 1989. 'Comment on "Are rain rate processes self-similar?" by B. Kedem and L. S. Chiu', *Wat. Resour. Res.*, **25**, 577–579.
Lovejoy, S. and Schertzer, D. 1991. 'Multifractal analysis techniques and rain and cloud fields from 10^{-3} to 10^{6} m' in Schertzer, D. and Lovejoy, S. (Eds), *Scaling, Fractals and Non-linear Variability in Geophysics*. Kluwer, Dordrecht. pp. 111–144.
Mackay, R. and Riley, M. S. 1991. 'The problem of scale in the modelling of groundwater flow and transport processes' in *Chemodynamics of Groundwaters, Proc. Workshop November 1991, Mont Sainte-Odile, France.* EAWAG, EERO, PIR "Environment" of CNRS, IMF Université Louis Pasteur Strasbourg. pp. 17–51.
Mandelbrot, B. B. 1967. 'How long is the coast of Britain? Statistical self-similarity and fractional dimension', *Science*, **156**, 636–638.
Mandelbrot, B. B. 1977. *Fractals: Form, Chance, and Dimension*. Freeman, San Francisco. 365 pp.

Mandelbrot, B. B. and Van Ness, J. W. 1968. 'Fractional Brownian motions, fractional noises and applications', *SIAM Rev.*, **10**, 422–437.
Mandelbrot, B. B. and Wallis, J. R. 1968. 'Noah, Joseph, and operational hydrology', *Wat. Resour. Res.*, **4**, 909–918.
Mandelbrot, B. B. and Wallis, J. R. 1969. 'Robustness of the rescaled range R/S in the measurement of noncyclic long run statistical dependence', *Wat. Resour. Res.*, **5**, 967–988.
Mantoglou, A. and Gelhar, L. W. 1987. 'Capillary tension head variance, mean soil moisture content, and effective specific soil moisture capacity of transient unsaturated flow in stratified soils', *Wat. Resour. Res.*, **23**, 47–56.
Marani, A., Rigon, R., and Rinaldo, A. 1991. 'A note on fractal channel networks', *Wat. Resour. Res.*, **27**, 3041–3049.
Matheron, G. 1967. 'Eléments pour une théorie des milieux poreux', cited in de Marsily (1986).
Matheron, G. 1973. 'The intrinsic random functions and their applications', *Adv. Appl. Prob.*, **5**, 438–468.
Mazac, O., Kelly, W. E., and Landa, I. 1985. 'A hydrogeophysical model for relations between electrical and hydraulic properties of aquifers', *J. Hydrol.*, **85**, 1–19.
McKenzie, N. J. and Austin, M. P. 1993. 'A quantitative Australian approach to medium and small scale surveys based on soil stratigraphy and environmental correlation', *Geoderma*, **57**, 329–355.
McKenzie, N. J. and MacLeod, D. A. 1989. 'Relationships between soil morphology and soil properties relevant to irrigated and dryland agriculture', *Aust. J. Soil. Res.*, **27**, 235–258.
McKerchar, A. I. and Pearson, C. P. 1989. 'Flood frequency in New Zealand', *Publ. No. 20*, Hydrology Centre, Christchurch, 87 pp.
Mein, R. 1993. 'Flood hydrology', *Catchword, May 1993*, Cooperative Centre for Catchment Hydrology, Monash University, Clayton, Victoria.
Menevau, C. and Sreenivasan, K. R. 1987. 'Simple multifractal cascade model for fully developed turbulence', *Phys. Rev. Lett.*, **59**, 1424–1427.
Menevau C., Sreenivasan, K. R., Kailasnath, P., and Fan, M. S. 1990. 'Joint multifractal measures: theory and applications to turbulence', *Phys. Rev. A*, **41**, 894–913.
Mesa, O. J. and Mifflin, E. R. 1986. 'On the relative role of hillslope and network geometry in hydrologic response' in Gupta, V. K., Rodríguez-Iturbe, I., and Wood, E. F. (Eds), *Scale Problems in Hydrology*. D. Reidel, Dordrecht, pp. 1–18.
Mesa, O. J. and Poveda, G. 1993. 'The Hurst effect: the scale of fluctuations approach', *Wat. Resour. Res.*, **29**, 3995–4002.
Miall, A. D. 1985. 'Architectural-element analysis: a new method of facies analysis applied to fluvial deposits', *Earth-Sci. Rev.*, **22**, 261–308.
Miller, E. E. 1990. 'Scaling in soil physics — introduction' in Hillel, D. and Elrick, D. E. (Eds), *Scaling in Soil Physics: Principles and Applications*. Vol. 25. Soil Science Society of America. pp. xvii–xxi.
Miller, E. E. and Miller, R. D. 1956. 'Physical theory of capillary flow phenomena', *J. Appl. Phys.*, **27**, 324–332.
Milly, P. C. and Eagleson, P. S. 1987. 'Effects of spatial variability on annual average water balance', *Wat. Resour. Res.*, **23**, 2135–2143.
Milne, G. 1935. 'Some suggested units of classification and mapping particularly for East African soils', *Soil. Res.*, Berlin, **4**, 183–198.
Minshall, N. E. 1960. 'Predicting storm runoff on small experimental watersheds', *J. Hydraul. Div. Am. Soc. Civ. Engin.*, **86**, (HY8) 17–38.
Montgomery, D. R. and Foufoula-Georgiou, E. 1993. 'Channel network source representation using digital elevation models', *Wat. Resour. Res.*, **29**, 3925–3934.
Moore, I. D. and Burch, G. J. 1986. 'Sediment transport capacity of sheet and rill flow: application of unit stream power theory', *Wat. Resour. Res.*, **22**, 1350–1360.
Moore, I. D. and Grayson, R. B. 1991. 'Terrain based prediction of runoff with vector elevation data', *Wat. Resour. Res.*, **27**, 1177–1191.
Moore, I. D., Burch, G. J., and Mackenzie, D. H. 1988. 'Topographic effects on the distribution of surface soil water and the location of ephemeral gullies', *Trans. Am. Soc. Agric. Engin.*, **31**, 1098–1107.
Moore, I. D., Grayson, R. B., and Ladson, A. R. 1991. 'Digital terrain modelling: a review of hydrological, geomorphological, and biological applications', *Hydrol. Process.*, **5**, 3–30.
Moore, I. D., Gessler, P. E., Nielsen, G. A., and Peterson, G. A. 1993a. 'Soil attribute prediction using terrain analysis', *Soil Sci. Soc. Am. J.*, **57**, 443–452.
Moore, I. D., Lewis, A., and Gallant, J. C. 1993b. 'Terrain attributes: estimation methods and scale effects' in Jakeman, A. J., Beck, M. B., and McAleer, M. (Eds), *Modelling Change in Environmental Systems*. Wiley, Chichester 189–214.
Moore, I. D., Turner, A. K., Wilson, J. P., Jenson, S. K., and Band, L. E., 1993c. 'GIS and land surface-subsurface process modeling' in Goodchild, M. F., Parks, B. O., and Steyaert, L. T. (Eds), *Geographic Information Systems and Environmental Modeling*. Oxford University Press, Oxford. pp. 196–230.
Morel-Seytoux, H. J. 1988. 'Soil–aquifer–stream interactions — a reductionist attempt toward physical-stochastic integration', *J. Hydrol.*, **102**, 355–379.
Neuman, S. P. 1990. 'Universal scaling of hydraulic conductivities and dispersivities in geologic media', *Wat. Resour. Res.*, **26**, 1749–1758.
Neuman, S. P. and Orr, S. 1993. 'Prediction of steady state flow in nonuniform geologic media by conditional moments: exact nonlocal formalism, effective conductivities, and weak approximations', *Wat. Resour. Res.*, **29**, 341–364.
National Research Council (NRC) 1988. *Estimating Probabilities of Extreme Floods: Methods and Recommended Research*. National Academy Press, Washington. 141 pp.
National Research Council (NRC) 1991. *Opportunities in the Hydrologic Sciences*. National Academy Press, Washington. 348 pp.
Obled, Ch. 1990. 'Hydrological modeling in regions of rugged relief' in Lang, H. and Musy, A. (Eds), *Hydrology in Mountainous Regions. I — Hydrological Measurements; the Water Cycle (Proc. Lausanne Symp., August 1990), IAHS Publ.*, **193**, 599–613.
Oboukhov, A. 1962. 'Some specific features of atmospheric turbulence', *J. Fluid Mech.*, **13**, 77–81.
O'Connell, P. E. 1991. 'A historical perspective' in Bowles, D. S. and O'Connell, P. E. (Eds), *Recent Advances in the Modeling of Hydrologic Systems*. Kluwer, Dordrecht. pp. 3–30.

Oliver, M. A. and Webster, R. 1986. 'Semi-variograms for modelling the spatial pattern of landform and soil properties', *Earth Surf. Process. Landforms*, **11**, 491–504.
O'Loughlin, E. M. 1981. 'Saturation regions in catchments and their relation to soil and topographic properties', *J. Hydrol.*, **53**, 229–246.
O'Loughlin, E. M. 1986. 'Prediction of surface saturation zones in natural catchments by topographic analysis', *Wat. Resour. Res.*, **22**, 794–804.
Orlanski, I. 1975. 'A rational subdivision of scales for atmospheric processes', *Bull. Am. Meteorol. Soc.*, **56**, 527–530.
Padmanabhan, G. and Rao, A. R. 1988. 'Maximum entropy spectral analysis of hydrologic data', *Wat. Resour. Res.*, **25**, 1519–1533.
Parsons, A. J., Abrahams, A. D., and Luk, S.-H. 1990. 'Hydraulics of interrill overland flow on a semi-arid hillslope, southern Arizona', *J. Hydrol.*, **117**, 255–273.
Pattison, A. 1965. 'Synthesis of hourly rainfall data', *Wat. Resour. Res.*, **1**, 489–498.
Pearce, A. J., Stewart, M. K., and Sklash, M. G. 1986. 'Storm runoff generation in humid headwater catchments. 1. Where does the water come from?', *Wat. Resour. Res.*, **22**, 1263–1272.
Pearson, C. P. 1993. 'Regional flood frequency analysis for small New Zealand basins. 2. Flood frequency groups', *J. Hydrol. (New Zealand)*, **30**, 77–92.
Philip, J. R. 1957. 'The theory of infiltration, 1–7', *Soil Sci.*, **83–85**.
Pilgrim, D. H. 1977. 'Isochrones of travel time and distribution of flood storage from a tracer study on a small watershed', *Wat. Resour. Res.*, **13**, 587–595.
Pilgrim, D. H. 1983 'Some problems in transferring hydrological relationships between small and large drainage basins and between regions', *J. Hydrol.*, **65**, 49–72.
Pilgrim, D. H. (Ed.) 1987. *Australian Rainfall and Runoff.* The Institution of Engineers, Barton, ACT. 374 pp.
Pilgrim, D. H. and Cordery, I. 1975. 'Rainfall temporal patterns for design floods', *J. Hydraul. Div. ASCE*, **101**, 81–95.
Raupach, M. R. and Finnigan, J. J. 1995. 'Scale issues in boundary layer — meteorology: surface energy balances in heterogenous terrain', *Hydrol. Process.*, **9**, 589–612.
Rawls, W. J., Brakensiek, D. L., and Miller, N. 1983. 'Green–Ampt infiltration parameters from soils data', *J. Hydraul. Engin.*, **109**, 62–70.
Reybold, W. U. and TeSelle, G. W. 1989. 'Soil geographic data bases', *J. Soil Wat. Conserv.*, Jan–Feb, 28–29.
Richards, K. 1982. *Rivers — Form and Process in Alluvial Channels.* Methuen, London. 361 pp.
Rinaldo, A., Marani, A., and Rigon, R. 1991. 'Geomorphologic dispersion', *Wat. Resour. Res.*, **27**, 513–525.
Robinson, J. S. and Sivapalan, M. 1995. 'Catchment-scale runoff generation model by aggregation and similarity analysis', *Hydrol. Process.*, **9**, 000–000.
Rodríguez-Iturbe, I. 1986. 'Scale problems in hydrologic processes' in Shen, H. W., Obeysekera, J. T. B., Yevjevich, V., and Decoursey, D. G. (Eds), *Multivariate Analysis of Hydrologic Processes.* Colorado State University. pp. 51–65.
Rodríguez-Iturbe, I. and Gupta, V. K. (Eds) 1983. 'Scale problems in hydrology', *J. Hydrol.*, **65** (spec. issue).
Rodríguez-Iturbe, I. and Valdés, J. B. 1979. 'The geomorphologic structure of hydrologic response', *Wat. Resour. Res.*, **15**, 1409–1420.
Rodríguez-Iturbe, I., Ijjász-Vásquez, E. J., Bras, R. L., and Tarboton, D. G. 1992a. 'Power law distributions of discharge mass and energy in river basins', *Wat. Resour. Res.*, **28**, 1089–1093.
Rodríguez-Iturbe, I., Rinaldo, A., Rigon, R., Bras, R. L., Marani, A., and Ijjász-Vásquez, E. 1992b. 'Energy dissipation, runoff production and the 3-dimensional structure of river basins', *Wat. Resour. Res.*, **28**, 1095–1103.
Rogers, A. D. 1992. 'The development of a simple infiltration capacity equation for spatially variable soils', *B.E. Thesis*, Univ. of West. Aust., Nedlands. 64 pp.
Rosso, R., Peano, A., Becchi, I., Bemporad, G. A. (Eds.) 1994. 'Advances in distributed hydrology' in *Proceedings of a Workshop held in Bergamo, Italy, June 1992.* Water Resources Publications, 416 pp.
Rosso, R., Bacchi, B., and La Barbera, P. 1991. 'Fractal relation of mainstream length to catchment area in river networks', *Wat. Resour. Res.*, **27**, 381–387.
Rubin, Y. and Gómez-Hernández, J. J. 1990. 'A stochastic approach to the problem of upscaling of conductivity in disordered media: theory and unconditional numerical simulations', *Wat. Resour. Res.*, **26**, 691–701.
Rubin, Y., Mavko, G., and Harris, J. 1992. 'Mapping permeability in heterogeneous aquifers using hydrologic and seismic data', *Wat. Resour. Res.*, **28**, 1809–1816.
Russo, D. and Jury, W. A. 1987. 'A theoretical study of the estimation of the correlation scale in spatially variable fields 1. stationary fields', *Wat. Resour. Res.*, **23**, 1257–1268.
Russo, D. 1992. 'Upscaling of hydraulic conductivity in partially saturated heterogeneous porous formation', *Wat. Resour. Res.*, **28**, 397–409.
Schönwiese, C. D. 1979. *Klimaschwankungen.* Springer, Berlin.
Schumm, S. A. 1956. 'Evolution of drainage systems and slopes in Badlands at Perth Amboy, New Jersey, *Geol. Soc. Am. Bull.*, **67**, 597–646.
Sellers, P. J., Hall, F. G., Asrar, G., Strebel, D. E., and Murphy, R. E. 1992. 'An overview of the First International Satellite Land Surface Climatology Project (ISLSCP) Field Experiment (FIFE)', *J. Geophys. Res.*, **97**(D17), 18345–18371.
Shreve, R. L. 1966. 'Statistical law of stream numbers', *J. Geol.*, **74**, 17–37.
Shreve, R. L. 1967. 'Infinite topologically random networks', *J. Geol.*, **75**, 178–186.
Silliman, S. E. and Wright, A. L. 1988. 'Stochastic analysis of paths of high hydraulic conductivity in porous media', *Wat. Resour. Res.*, **24**, 1901–1910.
Sivapalan, M. 1993. 'Linking hydrologic parameterisations across a range of scales: hillslope to catchment to region' in Bolle, H.-J., Feddes, R. A., and Kalma, J. D. (Eds), *Exchange Processes at the Land Surface for a Range of Space and Time scales, IAHS Publ.*, **212**, 115–123.

Sivapalan, M. and Viney, N. R. 1994a. 'Large scale catchment modelling to predict the effects of land use and climate', *Water, J. Aust. Wat. Wastewat. Assoc.*, **21,** 33–37.

Sivapalan, M. and Viney, N. R. 1994b. 'Application of a nested catchment model for predicting the effects of changes in forest cover'. In: Ohta, T. (Ed.) *Proc. Int. Symp. Forest Hydrology*, Univ. Tokyo, Tokyo. IUFRO, pp. 315–322.

Sivapalan, M. and Wood, E. F. 1986. 'Spatial heterogeneity and scale in the infiltration response of catchments' in Gupta, V. K., Rodriguez-Iturbe, I., and Wood, E. F. (Eds), *Scale Problems in Hydrology*. D. Reidel, Dordrecht. pp. 81–106.

Sivapalan, M., Beven, K., and Wood, E. F. 1987. 'On hydrologic similarity, 2. A scaled model of storm runoff production', *Wat. Resour. Res.*, **23,** 2266–2278.

Sivapalan, M., Wood, E. F., and Beven, K. J. 1990. 'On hydrologic similarity. 3. A dimensionless flood frequency model using a generalized geomorphologic unit hydrograph and partial area runoff generation', *Wat. Resour. Res.*, **26,** 43–58.

Skoda, G. 1987. 'Fractal dimension of rainbands over hilly terrain', *Meteorol. Atmos. Phys.*, **36,** 74–82.

Smagorinsky, J. 1974. 'Global atmospheric modeling and the numerical simulation of climate' in Hess, W. N. (Ed.), *Weather and Climate Modification*. Wiley, New York. pp. 633–686.

Smith, J. A. 1992. 'Representation of basin scale in flood peak distribution', *Wat. Resour. Res.*, **28,** 2993–2999.

Snell, J. D. and Sivapalan, M. 1994a. 'On geomorphological dispersion in natural catchments and the geomorphological unit hydrograph', *Wat. Resour. Res.*, **30,** 2311–2323.

Snell, J. D., and Sivapalan, M. 1994b. 'Threshold effects in geomorphological parameters extracted from DEM's', *Trans. Jpn Geomorph. Union*, **15A,** 67–93.

Snell, J. D. and Sivapalan, M. 1995. 'Application of the meta-channel concept: Construction of the meta-channel hydraulic geometry for a natural catchment', *Hydrol. Process.*, **9,** 485–506.

Sreenivasan, K. R. 1991. 'Fractals and multifractals in fluid turbulence', *Annu. Rev. Fluid Mech.*, **23,** 539–600.

Stommel, H. 1963. 'Varieties of oceanographic experience', *Science*, **139,** 572-576.

Strahler, A. N. 1957. 'Quantitative analysis of watershed geomorphology', *Trans. Am. Geophys. Union*, **38,** 913–920.

Strahler, A. N. 1964. 'Quantitative geomorphology of drainage basins and channel networks' in Chow, V. T. (Ed.), *Handbook of Applied Hydrology*. McGraw-Hill, New York. pp. 4.39–4.76.

Stull, R. B. 1988. *An Introduction to Boundary Layer Meteorology*. Kluwer Academic, Dordrecht. 666 pp.

Surkan, A. J. 1969. 'Synthetic hydrographs: effects of network geometry', *Wat. Resour. Res.*, **5,** 112–128.

Tabios, G. Q. and Salas, J. D. 1985. 'A comparative analysis of techniques for spatial interpolation of precipitation', *Wat. Resour. Bull.*, **21,** 365–380.

Tarboton, D. G., Bras, R. L., and Rodriguez-Iturbe, I. 1988. 'The fractal nature of river networks', *Wat. Resour. Res.*, **24,** 1317–1322.

Tarboton, D. G., Bras, R. L., and Rodriguez-Iturbe, I. 1989. 'Scaling and elevation in river networks', *Wat. Resour. Res.*, **25,** 2037–2051.

Tarboton, D. G., Bras, R. L., and Rodríguez-Iturbe, I. 1991. 'On the extraction of channel networks from digital elevation data' in Beven, K. J. and Moore, I. D. (Eds), *Terrain Analysis and Distributed Modelling in Hydrology*. Wiley, Chichester. pp. 85–106.

Thiessen, A. H. 1911. 'Precipitation averages for large areas', *Monthly Weather Rev.*, **39,** 1082–1084.

Tillotson, P. M. and Nielsen, D. R. 1984. 'Scale factors in soil science', *Soil Sci. Soc. Am. J.*, **48,** 953–959.

Troch, P. A., De Troch, F. P., Mancini, M., and Wood, E. F. 1995. 'Stream network morphology and storm response in humid catchments', *Hydrol. Process.*, **9,** (at press).

Troutman, B. M. and Karlinger, M. R. 1985. 'Unit hydrograph approximations assuming linear flow through topologically random networks', *Wat. Resour. Res.*, **21,** 743–754.

USDA-SCS (US Department of Agriculture Soil Conservation Service) 1986. 'Urban hydrology for small watersheds', *Tech. release 55*, USDA-SCS, Washington.

van der Tak, L. D. and Bras, R. L. 1990. 'Incorporating hillslope effects into the geomorphologic instantaneous unit hydrograph', *Wat. Resour. Res.*, **26,** 2393–2400.

Vertessy, R. and Band, L. 1993. 'Cross comparison of terrain analysis methods for hydrologic simulations' in Kalma, J., Sivapalan, M., and Wood, E. (Eds), *Scale Issues in Hydrological/Environmental Modelling, Proc. Workshop Robertson, CRES*. Australian National University, Canberra. p. 16.

von Bandat, H. F. 1962. *Aerogeology*. Gulf Publishing, Houston. 350 pp.

Voss, R. F. 1985. 'Random fractals: characterisation and measurement' in Pynn, R. and Skjeltorp, A. (Eds), *Scaling Phenomena in Disordered Systems*. Plenum Press, New York. pp. 1–11.

Voss, R. F. 1988. 'Fractals in nature: from characterization to simulation' in Pietgen, H. and Saupe, D. (Eds), *The Science of Fractal Images*. Springer, New York. pp. 21–70.

Walker, R. G. (Ed.) 1984. *Facies Models*. Geological Association of Canada, Toronto. 317 pp.

Wallis, J. R. 1988. 'Catastrophes, computing and containment: living with our restless habitat', *Spec. Sci. Technol.*, **11,** 295–315.

Weinberg, G. M. 1975. *An Introduction to General Systems Thinking*. Wiley, New York. 279 pp.

Wenzel, H. G. 1982. 'Rainfall for urban stormwater design' in Kibler, D. F. (Ed.), *Urban Storm Water Hydrology. Wat. Resour. Monogr. 7*. American Geophysical Union, Washington, pp. 35–64.

White, I. 1988. 'Measurement of soil physical properties in the field' in Steffen, W. L. and Denmead, O. T. (Eds), *Flow and Transport in the Natural Environment: Advances and Applications*. Springer, Berlin. pp. 59–85.

Willgoose, G. and Kuczera, G. 1995. 'Estimation of sub-grid scale kinematic wave parameters for hillslopes', *Hydrol. Process.*, **9,** 000–000.

Willgoose, G. and Riley, S. 1993. 'Scale dependence of runoff and the hydrology of a proposed mine rehabilitation' in Institution of Engineers (Ed.), *Towards the 21st Century, Hydrology and Water Resources Symposium, Newcastle, 1993*. The Institution of Engineers, Barton, ACT. pp. 159–164.

Willgoose, G. R., Bras, R. L., and Rodríguez-Iturbe, I. 1991. 'A physically based coupled network growth and hillslope evolution model: 1. Theory', *Wat. Resour. Res.*, **27,** 1671–1684.

Williams, R. E. 1988. 'Comment on "Statistical theory of groundwater flow and transport: pore to laboratory, laboratory to formation, and formation to regional scale" by Gedeon Dagan', *Wat. Resour. Res.*, **24,** 1197–1200.

Williams, J., Ross, P., and Bristow, K. 1992. 'Prediction of the Campbell water retention function from texture, structure, and organic matter' in Van Genuchten, M. Th., Leij, F. J., and Lund, L. J. (Eds.), *Proc. Int. Workshop on Indirect Methods for Estimating the Hydraulic Properties of Unsaturated Soils*. Univ. California, Riverside 427–441.
Wiltshire, S. E. 1985. 'Grouping basins for regional frequency analysis', *Hydrol. Sci. J.*, **30,** 151–159.
World Meteorological Organisation (WMO) 1969. 'Manual for depth–area–duration analysis of storm precipitation', *WMO No. 237, Tech. Pap.*, **129,** 1–31.
Woo, M.-K., Heron, R., Marsh, P., and Steer, P. 1983a. 'Comparison of weather station snowfall with winter snow accumulation in High Arctic basins', *Atmosphere–Ocean*, **21,** 312–325.
Woo, M.-K., Marsh, P., and Steer, P. 1983b. 'Basin water balance in a continuous permafrost environment' in *Permafrost, Fourth International Conference*. National Academy Press, Washington. pp. 1407–1411.
Wood, E. F. and Hebson, C. S. 1986. 'On hydrologic similarity. 1. Derivation of the dimensionless flood frequency curve', *Wat. Resour. Res.*, **22,** 1549–1554.
Wood, E. F., Sivapalan, M., Beven, K., and Band, L. 1988. 'Effects of spatial variability and scale with implications to hydrologic modeling', *J. Hydrol.*, **102,** 29–47.
Wood, E. F., Sivapalan, M., and Beven, K. 1990. 'Similarity and scale in catchment storm response', *Rev. Geophys.*, **28,** 1–18.
Woods, R. A., Sivapalan, M., and Duncan, M. J. 1995. 'Investigating the representative elementary area concept: an approach based on field data', *Hydrol. Process.*, **9,** 000–000.
Woolhiser, D. A. and Goodrich, D. 1988. 'Effect of storm rainfall intensity patterns on surface runoff', *J. Hydrol.*, **102,** 335–354.
Woolhiser, D. A. and Osborne, H. B. 1985. 'A stochastic model of dimensionless thunderstorm rainfall', *Wat. Resour. Res.*, **21,** 511–522.
Wu, Y.-H., Yevjevich, V., and Woolhiser, D. A. 1978. 'Effects of surface roughness and its spatial distribution on runoff hydrographs', *Colo. State Univ., Fort Collins, Colo., Hydrol. Pap.*, **96,** 47 pp.
Wu, Y.-H., Woolhiser, D. A., and Yevjevich, V. 1982. 'Effects of spatial variability of hydraulic resistance of runoff hydrographs', *J. Hydrol.*, **59,** 231–248.
Yang, C. T. 1976. 'Minimum unit stream power and fluvial hydraulics', *J. Hydraul. Div. ASCE*, **102**(HY7), 919–934.
Yang, C. T. and Song, C. C. S. 1979. 'Theory of minimum rate of energy dissipation', *J. Hydraul. Div. ASCE*, **105**(HY7), 769–784.
Yeh, T. C. J., Gelhar, L. W., and Gutjahr, A. L. 1985. 'Stochastic analysis of unsaturated flow in heterogeneous soils. 1. Statistically isotropic media', *Wat. Resour. Res.*, **21,** 447–456.
Yevjevich, V. 1972. *Stochastic Processes in Hydrology*. Water Resources Publications, Fort Collins. 276 pp.
Zuidema, P. K. 1985. *Hydraulik der Abflussbildung während Starkniederschlägen*. Mitt. VAW 79, ETH Zürich, 150 pp.