31 Parsimonious spatial representation of tropical soils within dynamic rainfall–runoff models

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INTRODUCTION

Models are used increasingly to simulate hydrological processes within tropical regions. There is now a wealth of publications addressing evaporation modelling (particularly wet-canopy evaporation) of local areas of tropical forest in, for example, Niger (Gash *et al*., 1997), Guyana (Jetten, 1996), Puerto Rico (Schellekens *et al*., 1999), Columbia (Marin *et al*., 2000) and Indonesia (Asdak *et al*., 1999; van Dijk and Bruijnzeel, 2001). Elsewhere in this volume, Roberts *et al.* provide an overview of evaporation processes and modelling. Other modelling studies have addressed the impact of such tropical evaporation on regional climates and global circulation (e.g. Polcher and Laval, 1994; Zeng, 1999; Zeng and Neelin, 1999; Zhang *et al*., 2001). New studies using time-series models are highlighting the effects of cycles in the rainfall, such as the El Niño Southern Oscillation (ENSO) on tropical evaporation, riverflow and water quality (e.g. Zeng, 1999; Chappell *et al*., 2001; Krishnaswamy *et al*., 2001; Whitaker *et al*., 2001; Chappell, Tych *et al*., this volume). Similarly, models that simulate the generation of riverflow from the rainfall received by a tropical catchment are also beginning to be applied more frequently. These models include:

(1) Metric-conceptual models of waterflow, based upon transfer functions.[1]* For example, application of the DBM modelling approach to a nested catchment system in Malaysian Borneo (Chappell *et al*., 1999a) and the application of IHACRES to a large Thai basin (Scoccimarro *et al.*, 1999).
(2) Conceptual models of waterflow based upon stores and predetermined empirical relationships. For example, application of the Nash model to Kenyan catchments (Onyando and Sharma, 1995), the Modhydrolog model to a tropical catchment (Chiew *et al*., 1996), the Reservoir-Water-Balance-Simulation model to Namibian catchments (Hughes and Metzler, 1998), and the HBV-96 model (discussed in Barnes and Bonell, this volume) to catchments in Zimbabwe, Tanzania and Bolivia (Liden and Harlin, 2000).
(3) Conceptual models of waterflow incorporating spatially distributed, topographic information. For example, application of TOPMODEL in Cote d'Ivoire (Quinn *et al*., 1991), French Guiana (Molicova *et al*., 1997) and Malaysian Borneo (Chappell *et al*., 1998) and TOPOG in Peru (Vertessy and Elsenbeer, 1999) and Puerto Rico (Schellekens, 2000).
(4) Hydrochemical mixing models for water-path identification. For example, analysis of natural chemical signals within catchment waters of Queensland, Australia (Elsenbeer *et al*., 1995) and Tanzania (Sandstrom, 1996) and environmental isotopes also in Queensland, Australia (Barnes and Bonell, 1996).
(5) Hydrological models based on Geographic Information System (GIS) mapping. For example, use of remote sensing and other GIS data in runoff prediction in West Africa (Schultz,

1 Technical words with asterisks are detailed in Appendix 31.1 at the end of the chapter.

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1994) and the ANSWERS model in Queensland, Australia (Connolly *et al*., 1997).

(6) Process-based catchment models solving the Richards Equation (Richards, 1931). For example, application of the Système Hydrologique Européen (SHE) model in India (Refsgaard *et al*., 1992; Singh *et al*., 1999) and Zimbabwe (Refsgaard and Knudsen, 1996).

One of the key questions to be addressed when deciding which of these modelling approaches should be applied at a new location, is how complex does the model structure need to be to describe adequately the rainfall-runoff behaviour of a tropical catchment (and possibly also the behaviour of certain internal characteristics, such as saturation extent)? This becomes critical when it is appreciated that very little information is contained within most time-series of riverflow, so that *very simple models are often sufficient to forecast the rainfall-runoff behaviour of catchments* (Kirkby, 1975; Jakeman and Hornberger, 1993; Beven, 2001a; Kokkonen and Jakeman, 2001; Young, 2001; see also Barnes and Bonell, this volume), whether in tropical or temperate regions.

This can be illustrated with the application of a Data-Based-Mechanistic (DBM*) model to the rainfall-runoff behaviour of the 0.44 km^2 Baru Experimental Catchment in Sabah, Malaysian Borneo (cf. Chappell *et al*., 1999a). The structure of the model can be described in transfer-function* form:

$$q(k) = \frac{P}{1 - \Re z^{-1}} r_{\mathrm{eff}}(k - \delta) \tag{31.1}$$

where

$$r_{\mathrm{eff}}(k) = r(k)\left[\theta_{k-1} + \frac{1}{\tau_\theta}\{r - \theta_{k-1}\}\right] \tag{31.2}$$

and $q(k)$ is the riverflow at the time index k, $\Re$ is the recession or lag parameter, P is the system production or gain parameter, z^{-1} is the backward shift operator (i.e., $z^{-i}\, r(k) = r(k-i)$) which allows expansion to higher-order models, δ is the pure time delay to the initial response, and r is the catchment-average rainfall input, r_{eff} is the transformed input or the waterflow after the catchment non-linearity has been characterised, term θ_{k-1} is the linear component of the internal flow at the previous time-step, and τ_θ is the time constant (or residence time) of the non-linear component of the catchment behaviour (Young, 1984; Whitehead *et al*., 1979). Further explanation of the DBM approach is given in Chappell *et al*. (in this volume).

Where the pure time delay is zero, then State Dependent Parameter (SDP) identification (Young, 2001) may identify as few as three catchment behaviour parameters that capture the rainfall-runoff behaviour of a catchment. These parameters are: the basin lag or recession, the production or water-balance term, and the term describing the form of the catchment non-linearity (i.e., $\Re$, P, and τ_θ). To illustrate this point, the DBM approach is applied to a one-year record of hourly rainfall and riverflow data for the Baru

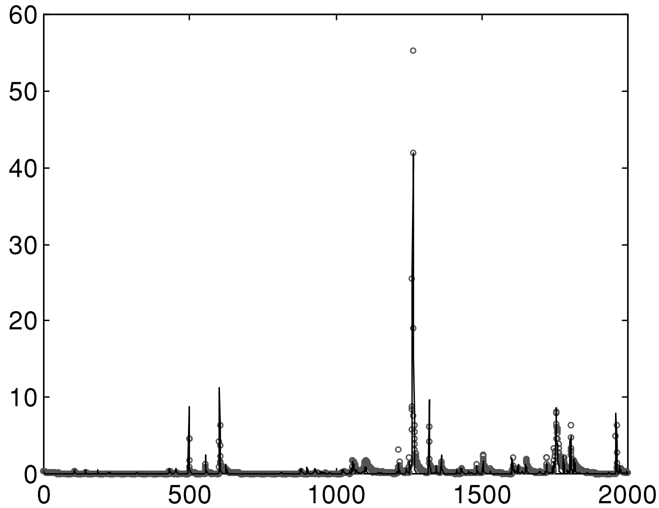

Figure 31.1 The results of a DBM model, incorporating the Bedford-Ouse Sub-Model to characterise the catchment non-linearity, applied to 1 year of hourly rainfall and riverflow data for the Baru Experimental Catchment, Sabah, Malaysian Borneo. The model has a Nash and Sutcliffe (1970) efficiency of 0.876 and a YIC of −9.93. The abscissa in time in hours, and the y-ordinate is riverflow in mm hr^{-1}. Observed riverflow is shown with a dotted line, and the modelled riverflow a solid line.

Experimental Catchment. A model efficiency (i.e. one minus the ratio of variance of the model errors to the variance of the observed data, expressed as a per cent) of 88% was achieved (Figure 31.1), indicating that the model captured most of the key dynamics inherent in the relationship between the incoming rainfall and outgoing riverflow. Such a three-parameter model indicates that one pathway dominates the catchment behaviour. DBM models do, however, allow for more complicated structures, i.e. multiple flow pathways. An objective statistic known as the 'Young Information Criterion' (YIC) can be used to examine if multiple pathways are observable within the dynamics of the catchment under study (Young, 2001). In the case of the year-long Baru Catchment dataset, however, the YIC did not support the use of more than one *dominant pathway* to route rainfall to the river.

Other studies have shown that complex (often physics-based) models do little better at forecasting riverflow time-series (in a split-sample validation) when compared with models with simple structures (and hence requiring few parameter values, i.e. parsimonious) e.g. Loague and Freeze, 1985; Franchini and Pacciani, 1991; Michaud and Sorooshian, 1994). In the tropical context, for example, the study of Refsgaard and Knudsen (1996) demonstrated that there was little additional forecasting benefit from applying the complex, physics-based MIKE-SHE model when compared to the NAM and WATBAL 'bucket models' to rainfall-runoff data from catchments in tropical Zimbabwe.

Potential value of physics-based modelling

Physics-based catchment models such as MIKE-SHE usually divide the model catchment into approximately 100 to 1000 landscape units. Over this distribution of landscape units, they then solve: (1) the momentum equation for waterflow (e.g. Darcy–Buckingham Equation for the subsurface flow: Darcy, 1856; Buckingham, 1907), and (2) the continuity equation.

These process-descriptions (or physical theories) incorporate soil-topographic parameters that (in principle) can be measured independently of the catchment's rainfall-runoff behaviour. As it may be possible to see the impact of a particular land-use activity on each of these soil-topographic parameters (e.g. soil permeability, porosity), it is then assumed that the impacts of land-use change on rainfall-runoff behaviour can be forecast via modifications to the model's parameters. This would suggest that such models have a significant advantage over 'bucket' and transfer-function models in their ability to predict the effects of internal catchment (e.g. land-use) changes on rainfall-riverflow behaviour. Physics-based models would, therefore, seem to have considerable value in the assessment of forestry, agricultural or urban impacts on the hydrological behaviour of tropical catchments.

Limitations to the testing and hence reliability of physics-based models

Currently, the physical algorithms on which many of these catchment-scale models are based were derived from observations of small-scale (10^{-1} m) phenomena and not on theory developed at the scale of the terrain-elements into which the modelled catchment has been divided (i.e. the model grid-scale*). These model grid elements are essentially the size of whole hillslopes, and tracer studies that strongly support Richards' formulations of waterflow in variably-saturated soil at this scale arguably do not exist (see e.g. Sherlock *et al*., 1995; Sherlock, 1997). As the rainfall-runoff behaviour of a catchment or indeed a single hillslope is a non-linear system, one cannot assume that a Constant of Proportionality such as soil permeability which represents the lumped behaviour of a small core, will be appropriate at the scale of a highly heterogeneous hillslope (Beven, 2001b). Validation of the model's assumed 'grid-scale physics' (and hence the parameter estimates indicated by model calibration) is clearly necessary. This testing process has, however, been hampered primarily by the lack of available techniques to measure the 'same' lumped grid/hillslope scale parameters independently in the field (Beven, 2001a). This problem is then compounded by:

(1) The perceived need to represent all of the profile and catenal changes in soil-rock permeability within a catchment (Chappell and Ternan, 1992) which has resulted in the development of physics-based models that require, for example, the specification of numerous permeability values over the modelled catchment.
(2) At the small-scale, some of the most sensitive* soil and topographic parameters (e.g. soil permeability), have a very high degree of unstructured spatial variability (Bonell *et al*., 1983; Elsenbeer and Cassel., 1993; Bonell with Balek, 1993). This has made it easy for those undertaking the modelling to not question almost any estimate of grid-scale* parameters identified by model calibration as being 'within the measured range'.
(3) The difficulty of identifying model parameters by calibration (i.e. parameter inversion process*), is magnified by the relatively recent observation that *many 'sets of parameters', each with very different values of each parameter, will give acceptable simulations of the same rainfall-runoff time-series* (Freedman *et al*., 1998; Beven, 2001a; Thiemann *et al*., 2001). The greater the model complexity, the more interaction between modelled parameters can take place during the calibration, so the more different the resultant parameter sets become, and finally the wider (or more uncertain) is the range of each model parameter.

There is, therefore, considerable merit in attempting to *constrain the complexity of catchment models to reduce the number of parameter sets that give acceptable model predictions, and thereby constrain the likely range of each model-calibrated parameter*. This then allows a more realistic comparison of grid-scale model-parameters estimated by inversion with those derived from field measurements (and appropriately up-scaled). Modelling where the *objective is to limit complexity* so that each parameter can be more narrowly defined, is called parsimonious. The group of catchment models, known as 'topographically-based hydrological models' (e.g. TOPMODEL and TOPOG_SBM) have at their core a simple/parsimonious structure, though they too are sometimes extended to become complex, parameter-demanding models. Clearly, in the tropics where there is a dearth of soil data, models which require fewer parameters and less complex spatial distributions of each parameter could be helpful.

TOPOGRAPHICALLY BASED, DYNAMIC RAINFALL-RUNOFF MODELS

The two most widely used topographically-based, dynamic catchment models are TOPMODEL (Beven and Kirkby, 1979; Beven, 1997), and TOPOG_SBM / TOPOG_DYNAMIC (Vertessy *et al*., 1994; Vertessy and Elsenbeer, 1999). The TOPOG variants were derived originally from the WETZONE model of O'Loughlin (1986). These models might be seen to be based on two key structural components:

(1) an index describing the degree of profile-saturation distributed in plan throughout the catchment (known as the 'topographic index' in TOPMODEL and the 'wetness index' in TOPOG), and
(2) a momentum equation capable of capturing the non-linear relation between rainfall and the river discharge generated.

(1) SPATIAL MAPPING OF SOIL SATURATION

The topographic index (λ) within TOPMODEL is normally (cf. Ambroise *et al*., 1996) defined as:

$$\lambda = \ln\left(\frac{a}{\tan\beta}\right) \quad (31.3)$$

where a is the upslope contributing area (in plan) to a given profile (m^2), and $\tan\beta$ is the local slope angle at that profile. A greater λ indicates a greater likelihood of saturation and hence surface-flow generation. The topographic index components indicate first, that as the subsurface contributing area to a local soil-rock profile in the catchment increases, so does the likelihood of increasing relative saturation. Second, the index incorporates the Darcian assumption of subsurface flow being proportional to the hydraulic gradient, and this can be approximated by the tangent of the ground-surface slope ($\tan\beta$). Thus, a steeper topographic slope is expected to give a greater hydraulic gradient and, therefore, increase the 'drainage potential' of a local soil-rock profile and thereby reduce its level of saturation. The topographic index mapped over the 0.44 km^2 Baru catchment in equatorial Borneo (using a 20 × 20 m Digital Terrain Model, or DTM) is given in Figure 31.2. Like the 'wetness index', λ is an index of 'hydrological similarity', which means that *elements of the terrain with the same index value are expected to behave in a similar hydrological manner*.

Over the last 20 years numerous studies have attempted to test this index against field observations of the dynamic spatial patterns of:

(a) the extent of surface saturation (e.g. O'Loughlin, 1981; Barling *et al*., 1994),
(b) soil moisture content (e.g. Burt and Butcher, 1983; Chappell and Franks, 1996; Sulebak *et al*., 2000),
(c) capillary potential (Molicova *et al*., 1997), and
(d) water-table level within boreholes (e.g. Troch *et al*., 1993; Jordan, 1994; Moore and Thompson, 1996).

The level of agreement between the observed and predicted saturated extent has been mixed, though Chappell and Franks (1996) have demonstrated that saturated extent may be predicted better on some slopes within a single catchment than others, and this may indicate spatial differences in hydrological complexity of the studied catchment. Some improvements to saturated area estimation have been made by:

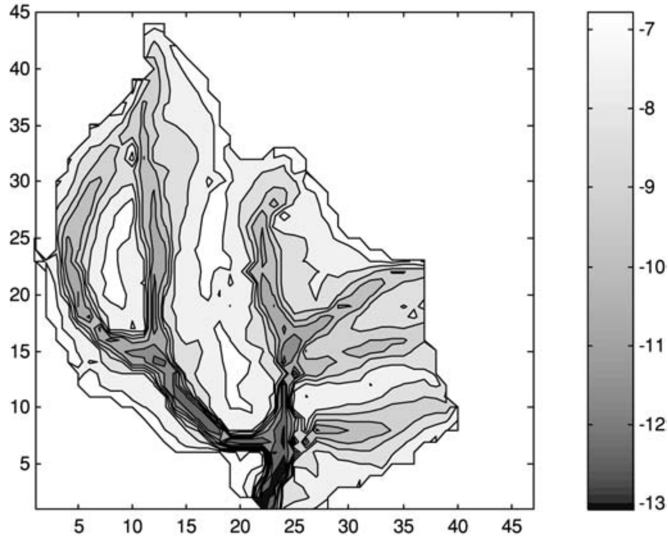

Figure 31.2 The map of the topographic index (λ) for the 0.44 km^2 Baru Experimental Catchment, Sabah, Malaysian Borneo. The index ranges from 5.99 to 12.99. The abscissa and y-ordinate axes show the number of grid cells. North is to the top of the figure.

(a) varying the model's transmissivity profile where borehole data are available (Lamb *et al*., 1997),
(b) development of Dynamic-TOPMODEL to reproduce varying subsurface contributing areas by allowing the index to vary (Beven and Freer, 2001), though allowing spatial variations in the transmissivity and index may add significantly to model complexity,
(c) use of the bedrock-surface, rather than ground-surface, to derive the topographic index (Freer *et al*., 1997), and
(d) evaluation of the effect of methods of analysing topography, and also the effects of different topographic grid sizes* on the values of transmissivity identified by TOPMODEL (Franchini *et al*., 1996; Saulnier *et al*., 1997; Brassington and Richards, 1998). Typically, these studies have shown that above a certain grid-size for the DTM (often 50 × 50 m), the effect on the catchment-average topographic index is to increase the T_0 term (i.e. lateral transmissivity, when the soil and weathered rock profile is saturated to the ground surface) within TOPMODEL calibrations. Clearly, this has implications for the interpretation of model-derived transmissivity and permeability values where coarse grids and parameter calibration routines are used.

While much effort has been directed towards the analysis of the topographic index, *comparatively little effort has been directed towards the second component of TOPMODEL (or indeed TOPOG), which contains the terms associated with the spatial distribution of soil permeability* (i.e. T_0 and m, where m is the exponential decay rate of lateral transmissivity with depth). Yet, soil permeability is invariably seen as the most important soil–rock property specified within complex physics-based models (Freeze, 1980; Rogers *et al*., 1985; Sherlock *et al*., 2000) and also in topographically-based models (Franchini *et al*., 1996).

(2) LATERAL PERMEABILITY DISTRIBUTION

Within TOPMODEL the river discharge (q_i) is generated by:

$$q_i = a_i r = T_0 e^{-\bar{S}_i/m} \tan\beta \tag{31.4}$$

where a_i is the upslope contributing area to a given point i per unit contour length (m^2), r is the catchment-average rainfall (m hr^{-1}), T_0 is the lateral transmissivity, when the soil and weathered rock profile is saturated to the ground-surface (m^2 hr^{-1}), $\bar{S}$ is the catchment-average, subsurface storage deficit (m), m is the exponential decay rate of lateral transmissivity with depth, and $\tan\beta$ is the tangent of the local slope angle at point i (m m^{-1}) (Quinn *et al*., 1991). The lateral transmissivity for varying saturated depths, T (m^2 hr^{-1}) is then:

$$T = T_0 \mathrm{e}^{-\bar{S}/m} \tag{31.5}$$

where the catchment-average saturation deficit is:

$$\bar{S} = -\frac{m}{A}\int \sum \zeta + \ln r(dA) \tag{31.6}$$

and A is the total catchment area (m^2) and ζ is the combined soil-topographic index. This can then be related to a catchment-average, water-table head using:

$$H = D_r - \left(\frac{\bar{S}}{\eta_{\mathrm{eff}}}\right) \tag{31.7}$$

where H is the water-table head above the solid rock (m), D_r is the depth to the solid rock (m), and η_{eff} is the effective porosity (m^3 m^{-3}) (Chappell *et al*., 1998). The lateral block permeability over the saturated part of the profile (K_{SM}, m hr^{-1}) is then:

$$K_{\mathrm{SM}} = \frac{T}{H} \tag{31.8}$$

It is clear that the vertical distribution of lateral permeability is described by a single exponential function (Eqn 31.5). The rate of decline of the exponential function is governed largely by the m parameter, which can be determined from the average hydrograph recession (i.e. Master Recession Curve) independent of rainfall-runoff calibration. Indeed, Lamb and Beven (1997) have developed an algorithm, 'MRC tool', to automate this task. Many river hydrographs, in both tropical and temperate catchments, can be described by an exponential recession component. So the application of a topographic model that assumes an exponential recession is a good starting point. What is perhaps 'fortuitous' is that many tropical catchments have soils where the permeability declines exponentially with depth (Figure 31.3; Beven, 1982; Sherlock, 1997). Often this is the result of a combination of the argillation process in the upper profile, combined with the effect of uninterrupted *in situ* chemical weathering of the lower layers (Fitzpatrick, 1971). A fuller description of the theoretical basis, assumptions and limitations of TOPMODEL is given in Kirkby (1975), Beven and Kirkby (1979), and Beven (1997), and on the internet site: http://www.es.lancs.ac.uk/hfdg/topmodel.html. Freely available model code and executables are also available on this internet site.

Clearly, soil permeability need not follow a monotonic exponential decline with depth but follow other monotonic relationships such as a power (Lancaster, 2000) or linear decline. In soils developed on deposits such as head and glacio-lacustrine drift in temperate environments or river terrace alluvium and volcanic tuff in temperate and tropical environments, then simple monotonic reductions of permeability with depth might not be observed (Chappell and Ternan, 1992, L. A. Bruijnzeel, pers. comm.). Non-monotonic changes of permeability with depth (e.g. where a C or R soil horizon is more permeable than the overlying B soil horizon) would be difficult to equate with TOPMODEL assumptions and also to reproduce within the parameterisation. Where soils exhibit non-monotonic declines in permeability, not only is model structure more difficult to define but so is the up-scaling of the point-scale, field measurements so that they are representative of the lateral permeabilities of whole soil profiles.

The point-scale measurements of soil permeability obtained with field tests such as ring permeametry (Bonell *et al*., 1983; Chappell and Ternan, 1997) cannot be compared directly to estimates of lateral permeability representative of the catchment-average soil profile or even the hillslope-average soil profile derived from model inversion* (Chappell *et al*., 1998). *The point-scale measurements of soil permeability need to be first 'up-scaled' to give estimates of lateral permeability of the catchment-average soil profile*. The permeabilities derived from up-scaling* are known as block permeabilities.* If we attempt to up-scale point-permeabilities for comparison with TOPMODEL-derived values, the resultant 'lateral block permeabilities' should be equal to something between an arithmetic and a harmonic mean of the core-scale values (Cardwell and Parsons, 1945). This uncertainty relates to the uncertain geometry of the flow pathways. It is often difficult to define a priori the up-scaling method more precisely (Wen and Gómez-Hernández, 1996).

To illustrate a methodology for evaluating soil representation within rainfall-runoff models and also highlight some of the difficulties associated with such a test, we will now summarise an approach described more fully in Chappell *et al*. (1998). This study examined a small tropical catchment, the Baru Catchment in Malaysian Borneo, comprised of Acrisol-Alisol (i.e.

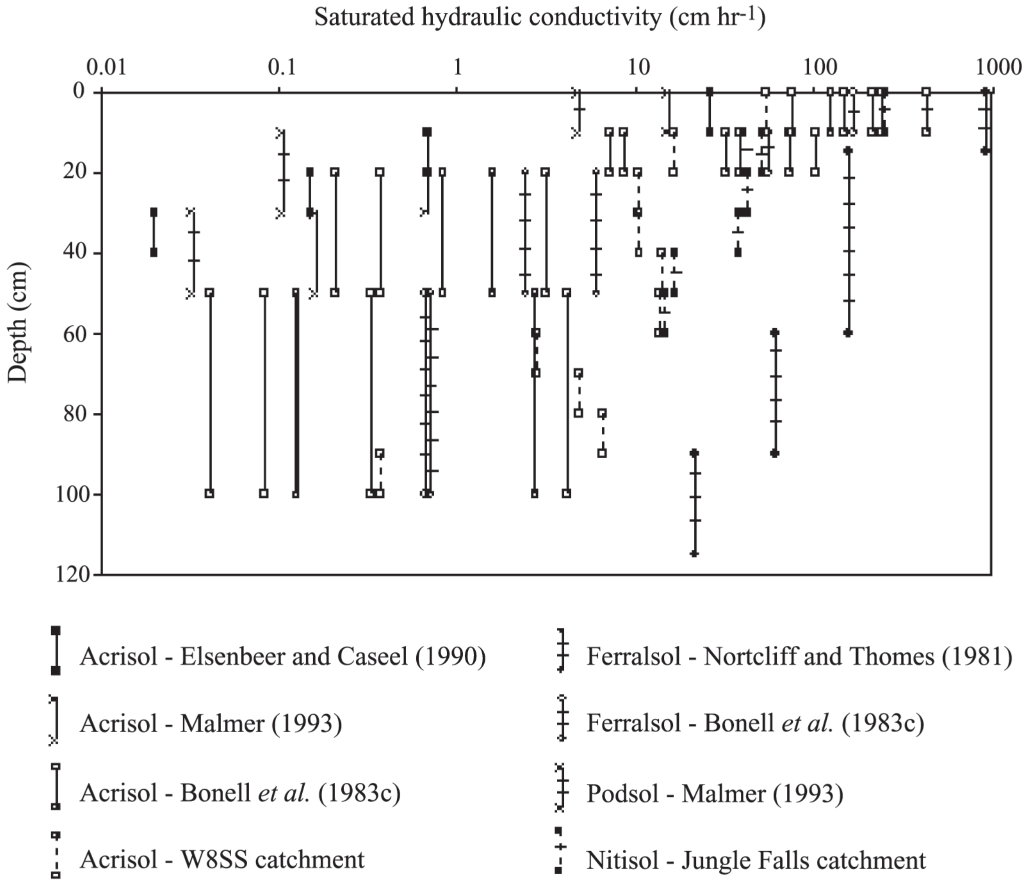

Figure 31.3 Profiles of saturated hydraulic conductivity (or 'soil permeability') from selected tropical soils. The vertical range that each published result represents is shown. (After Sherlock, 1997.)

USDA-Ultisol) soils (Chappell *et al.*, 1999b). This soil group is widespread throughout the tropics of SE Asia, West Africa and Amazonia (Bridges *et al.*, 1998).

EQUATING FIELD AND MODEL PERMEABILITY: A CASE STUDY OF A TROPICAL CATCHMENT

The Chappell *et al*. (1998) study was undertaken in natural forest (managed and disturbed) close to the Danum Valley Field Centre, Malaysian Borneo (5°01′N, 117°48.75′E), and had five key steps. These were: (1) field measurement of permeability, using relatively large soil cores, (2) up-scaling the core-based measurements to give lateral permeabilities representative of the whole hillslope soil profile, (3) deriving the catchment-average, lateral permeability profile – a procedure which is only realistic when a parsimonious model such as TOPMODEL is used, (4) comparison of up-scaled, field-derived permeabilities with model-derived permeabilities, and (5) derivation and preliminary testing of a new method for measuring the lateral permeability of the whole hillslope soil profiles.

(1) Point measurement of permeability

Ring permeametry (Bonell *et al.*, 1983; Chappell and Ternan, 1997) was used to measure the spatial distribution of soil permeability within a 12 km^2 region of only Acrisol-Alisol soils containing the Danum Valley Field Centre (Malaysian Borneo) and the catchment to be modelled. A total of 70 such permeability measurements on undisturbed, 30 cm diameter, 10 cm deep cores were taken from depths ranging from 0.05 to 3 m at crest-slope, side-slope and valley locations (Sinun, 1991; Chappell and Binley, 1992; Bidin, 1995; Sherlock, 1997; Chappell *et al.*, 1998). These data indicated that the spatial (vertical) variability in the permeability followed a simple monotonic decline with depth. The exponential function fitted to median permeability estimates for each depth, followed:

$$K_{S(med)} = 42e^{-2.188D}(R^2 = 0.92) \qquad (31.9)$$

where $K_{S(med)}$ is the core-scale, median permeability (m s^{-1}) at measurement depth D (m). This function fitted almost as well to the geometric mean permeability for each depth, i.e.

$$K_{S(geo)} = 52e^{-2.2907D}(R^2 = 0.88) \qquad (31.10)$$

where $K_{S(geo)}$ is the core-scale, geometric mean permeability (m s^{-1}) at measurement depth D (m). Clearly, these two models give some permeability at depths of many metres, but the local value will be several orders of magnitude less than that in the uppermost 3 metres of the soil-rock profile. Below 3 m, the soil-rock profile can, therefore, be assumed to have effectively no permeability.

(2) Up-scaling point measures of permeability

Within the study, an uncertain range of lateral block permeabilities were then estimated by using both arithmetic and harmonic averaging of the core-scale values. The calculated estimates of the lateral block permeability per the saturated part of the profile (K_{SC}) are presented in Figure 31.4. These estimates can be multiplied by the saturated depth to give the lateral transmissivity distribution. These data, unlike the original field measurements, could then be compared with results from the model inversion undertaken using catchment-scale data.

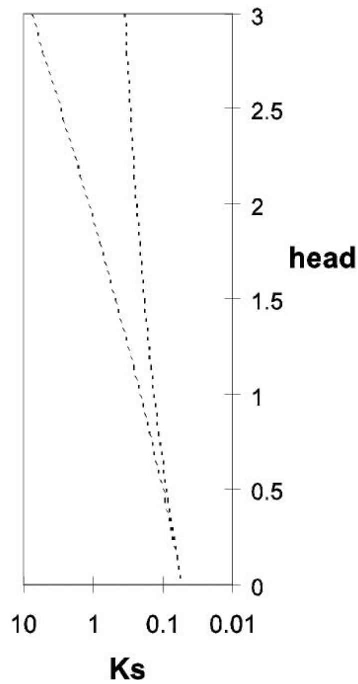

Figure 31.4 Lateral block permeability distribution derived by up-scaling core-based measurements from a 12 km^2 region around the Danum Valley Field Centre (Sabah, Malaysian Borneo) using a harmonic mean (right-hand curve) and an arithmetic mean (left-hand curve). The abscissa shows the lateral block permeability per saturated part of the soil profile (K_{SC}) in units of $\times$ 10^{-6} m s^{-1}. The *y*-ordinate shows the water table depth in metres, where zero is the rock-head (i.e. surface of the impermeable rock) and 3 m is the ground surface.

(3) Permeability estimation by catchment-model inversion

The 12 km^2 region in which the point permeability measurements were undertaken contains the 0.44 km^2 Baru Experimental Catchment. This catchment contains a distribution of data-logged raingauges and river gauging structures (Chappell *et al.*, 1999a). The Chappell *et al*. (1998) study used hourly rainfall and riverflow data lumped over the whole catchment for the period 6 May to 8 November 1995. Within this study a simple form of TOPMODEL requiring only four parameters (T_0, *m, Td* and SRMAX) was used to reduce the number of parameter sets that give acceptable efficiencies*, and hence help to constrain the range of individual parameters. The *Td* is the unsaturated zone time delay and SRMAX is the maximum root zone storage. Even with only four parameters, the range in parameters within physics-based and quasi-physics-based models can be so large as to include any field measurements. The Chappell *et al.* (1998) study, therefore, followed the approach of Franks *et al.* (1997). They demonstrated that where additional data are available for the catchment, such as a knowledge of the maximum and minimum extent of saturated areas, then this information could be used to reject model simulations that gave very unrealistic saturated extents. They further demonstrated that a large reduction in the uncertainty in the critical T_0 parameter could be obtained using this approach. More recently, Wooldridge *et al.* (2001) have similarly suggested that rejection of some parameter sets on the basis of inconsistency of simulations with additional internal state data (i.e. non-behavioural sets) constrains parameter uncertainty significantly.

As a result of the acknowledged effects of overly coarse DTM resolutions on the topographic index and thence the T_0 parameter (Franchini *et al.*, 1996; Saulnier *et al.*, 1997; Brassington and Richards, 1998), a fine resolution DTM (i.e. 10 $\times$ 10 m) was used. Ten thousand Monte Carlo simulations of TOPMODEL were run to identify the 'better' parameter sets. Field observations of the spatial extent of saturated soil profiles (Sinun, 1991; Sherlock, 1997; Chappell *et al.*, 1999b; P. Kukuon, pers. comm.) are were well within the range of 2 to 10% of the catchment during the hydrometeorological conditions experienced in 1995. The Monte Carlo results were, therefore, conditioned further by rejecting those model parameter sets that produced saturated areas either less than 2% or greater than 10% saturated area (i.e. parameter estimates associated with non-behavioural simulations

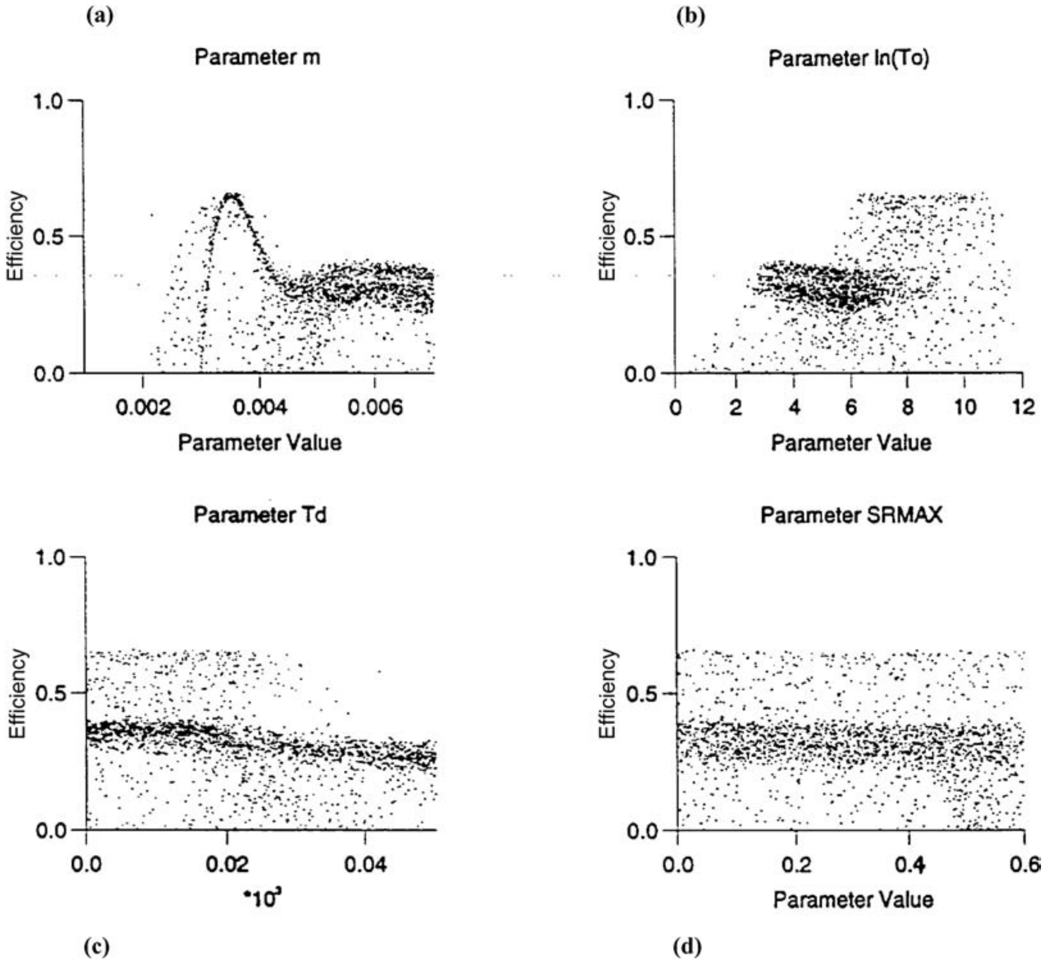

Figure 31.5 Scattergrams of model efficiency against (a) *m* (metres), (b) T_0 (ln[m^2 s^{-1}]), (c) *Td* ($\times$ 10^{-5} s m^{-1}), and (d) SRMAX (metres) parameters after comparison with discharge predictions that produce between 2% and 10% saturated area. (After Chappell *et al.*, 1998.)

were discarded). The result of the conditioning by saturated area extent was indeed seen to constrain the T_0 parameter (Figure 31.5), giving a plateau in the efficiency surface ranging from 0.189 to 4.910 m^2 s^{-1}. This range, combined with the peak value for *m* of approximately 3.5 $\times$ 10^{-3} m, was then used to calculate an uncertain range of lateral block permeability. This was presented over the range of water-tables that the model yielded, given the maximum and minimum T_0 values chosen (Figure 31.6). It is unrealistic to present lateral block permeabilities (or transmissivities) outside of the range of behavioural simulations (hence at the extremes of the exponential distribution). The predicted water-tables were indeed consistent with the patterns observed throughout the 12 km^2 Danum Valley study region (Bidin, 1995; Sherlock, 1997; P. Kukuon, pers. comm.). The mean predicted water-table head (above the effective rock-head of 3 m) during the simulation period was 1.24 m (Chappell *et al.*, 1998) and the lateral block permeability at this mean head was between 0.527 $\times$ 10^{-6} and 13.7 $\times$ 10^{-6} m s^{-1} from the modelling (K_{SM}).

(4) Comparison of model-derived permeabilities with up-scaled field values

Two key differences between the up-scaled, core-based permeability data and model-derived permeability data are observed in Figure 31.6:

(a) The range from the up-scaled core data (K_{SC}) of 0.158 $\times$ 10^{-6} to 0.311 $\times$ 10^{-6} m s^{-1} for the Baru Catchment soils is clearly much less than the model-derived range (Figure 31.6). This may indicate that: (i) some of the key conductive pathways (e.g., natural soil pipes*, percolines* or fractures) are not characterised adequately by core-scale measurements, or (ii) that the rainfall-riverflow response is not best characterised by permeabilities and hydraulic gradients but by more rapid response mechanisms such as the migration of a solitary wave across a hillslope water-table surface (Chappell *et al.*, 1998). A range of physics-based modelling studies (e.g., Sloan *et al.*, 1983; Blain and Milly, 1991) have similarly

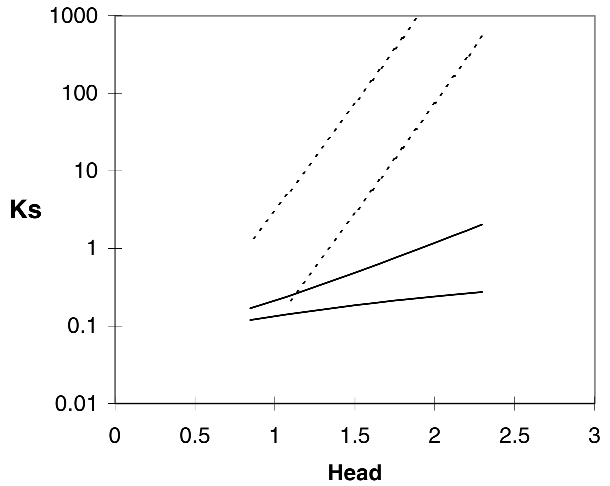

Figure 31.6 The uncertain band of lateral block permeability derived from the TOPMODEL inversion (between the broken lines, K_{SM}) and that derived by up-scaling core-based measurements (between the solid lines, K_{SC}). Lateral block permeability has units of $\times 10^{-6}$ m s^{-1} and water-table head is in metres. (After Chappell *et al*., 1998.)

indicated that effective permeabilities* derived by hillslope and catchment-scale inversion give higher estimates that those derived from point-scale, field measurements. Vertessy and Elsenbeer (1999) noted that their model-derived permeability estimates were much larger than their point estimates measured in an Amazonian catchment (and presumably larger than the lateral block permeabilities, if they had derived them). Bonell (1998), assessing current challenges of research on river generation processes, reported similar differences between field- and model- derived permeability estimates. The suggestion that models such as TOPMODEL may generate artificially high effective permeabilities as a result of its steady-state assumption (Wigmosta and Lettenmaier, 1999) does, however, need further investigation.

(b) The slope of the exponential distribution of the permeability band for up-scaled core measurements and model-derived measurements are significantly different (Figure 31.6), which may indicate that there is a greater non-linearity in the catchment response (here related largely to the subsurface system) than can be obtained from the core-scale measurements. Application of TOPOG_SBM to the generation of infiltration-excess overland flow within a small catchment in the headwaters of the Amazon similarly demonstrated that the measured permeability profile could not reproduce the degree of non-linearity observed in the rainfall-runoff response (Vertessy and Elsenbeer, 1999). Similar problems in trying to parameterise TOPOG_SBM and TOPOG_DYNAMIC with field data were found by Schellekens (2000) studying a small catchment in Puerto Rico (see Bonell, this volume). New versions of TOPMODEL using linear and power declines of permeability with depth rather than an exponential decline, have been developed recently (Ambroise *et al.*, 1996; Duan and Miller, 1997; Iorgulescu and Musy, 1997). These models have been applied successfully to catchments that do not have simple exponential recession curves (e.g. Lancaster, 2000). Given that the average rate of loss of permeability with depth is orders of magnitude greater with a catchment-scale inversion than that observed with point-scale measurements, it could be argued, therefore, that the poor correspondence of the field and model values are not attributable simply to the type of permeability function.

Given both of the differences highlighted, perhaps there is value in approaches that yield accurate hillslope-scale estimates of a 'mean' permeability, as these data might be compared with: (i) point-scale measures, up-scaled to the hillslope unit, and (ii) catchment-average permeability, derived from model inversion.

(5) Hillslope-scale permeability estimation

Chappell *et al*. (1998) developed a simplified methodology for estimating whole-hillslope permeability for an Acrisol-Alisol soil.

It is important to remember that the permeability of a soil and/or weathered rock is the 'rate of water-flow through a unit area of saturated media under a unit hydraulic gradient'. Therefore, if the propagation rate of a water pulse through a hillslope is monitored and then corrected for the gradient (and other effects such as lateral dispersion of a local source), then a very approximate estimate of the lateral block permeability can be obtained. Three further key details are needed:

(a) the distance between the stream and half way to the local catchment divide is assumed to be the approximate average travel distance of rainwater migrating through the subsurface system (i.e. pathline* length), at least for shallow groundwater catchments,
(b) by applying a steady-state pulse of water to the slope at this mid-point and tracing the pulse to the stream (using tensiometers and/or boreholes), we can obtain a distribution of propagation velocities (V) for travel to any down-slope location, and
(c) if we assume that the hydraulic gradient between the mid-slope and monitoring location can be approximated by the sine of the slope angle (i.e. the difference in gravitational potential over the slope distance) then lateral block permeability estimates can be derived from the propagation velocities.

Given the approximations within the method and the need for parametric simplicity, Chappell *et al*. (1998) estimated only two propagation velocities to each sampling point. The first velocity, V_S (m s^{-1}), is the ratio of: (a) the time from water injection to a steady-state response at a sampling location, and (b) the length of the pathline.* The second velocity, V_C (m s^{-1}), is the ratio of: (a) the centroid time between injection and steady-state response and (b) the length of the pathline. These values, assumed to be pore-water velocities, were multiplied by the effective porosity (η_{eff}) to estimate block permeabilities (discharges per unit area). An estimate of the η_{eff} (i.e., 0.025 m^3 m^{-3}) was derived from the difference between the total porosity and the moisture content at −1.5 kPa capillary potential on moisture release curves determined by Sherlock (1997).

During the pulse-wave tests of Sherlock (1997), Chappell *et al.* (1998) and Lancaster (2000), water is normally applied over a 1 m width normal to the slope. The width of the resultant plume is observed to expand with distance from the injection point. The propagation of the water, therefore, needs to be corrected for this dispersion by calculation of a dispersion factor, τ, which is the across-slope width of plume at the monitoring point, normalised by the width at the injection site.

The complete calculation of the estimate of the lateral block permeability derived from, for example V_S, is therefore

$$K_{SVS} = \frac{V_S \eta_{eff}}{\tau \sin \beta} \tag{31.11}$$

where K_{SVS} is the lateral block permeability between the mid-slope and down-slope monitoring point, as derived from V_s (m s^{-1}). Chappell *et al.* (1998) state that their aim was to present a tractable solution rather than one that includes all terms (after O'Loughlin, 1990).

Chappell *et al.* (1998) presented preliminary results of the application of this technique to the same tropical soil as that to which the catchment-scale model inversion was applied. The estimates of lateral block permeability derived from a single hillslope pulse-wave experiment (Reference name: TIKO) were K_{SVS} of 8.9 × 10^{-6} m s^{-1} and K_{SVS} 14.8 × 10^{-6} m s^{-1}. These results are not inconsistent with those derived from their model inversion (i.e. 0.527 × 10^{-6} and 13.7 × 10^{-6} m s^{-1}) (Figure 31.6). Further, the hillslope-derived estimates are clearly considerably larger than any of the up-scaled, core-based values (Figure 31.6). These results do not falsify the idea that the water plume applied to the experimental hillslope migrates under the influence of the natural soil pipes* and is less affected by the permeability distribution indicated by core-scale measurements. This same conclusion was reached by Lancaster (2000), who was able to undertake the same pulse-wave experiments, but replicated over several hillslopes within a 0.1 km^2 catchment in northwest England (UK). These preliminary results, therefore, support the idea that pulse-wave experiments may be a better method of deriving 'best estimates' of whole-hillslope permeabilities, and that they could be used to evaluate the estimates derived from catchment-scale modelling.

Clearly, derivation of a spatial distribution of whole-hillslope permeabilities do not give sufficient information on the form of the rainfall-runoff non-linearity to become the sole method of parameterising the soil-rock component of a catchment model. However, it is hoped that this may be a first step, with more conceptual work at the whole-hillslope scale being required. The central issue that arises from this case study is that core-scale measurements of key tropical soil parameters (notably soil-rock permeability), even after up-scaling, may not be meaningful for modelling catchment rainfall–runoff.

CONCLUSIONS

The objective of this chapter was to present a critical assessment of how soil-rock permeability is parameterised within dynamic rainfall-runoff models. As a result of the problem of 'parameter uncertainty'* arising from too complex a model structure, only a structurally-simple (i.e. parsimonious) model was thought capable of evaluation.

Four key conclusions arise that we believe are important to further developments in tropical catchment modelling:

(1) Further rainfall–runoff modelling in the tropics (and indeed elsewhere) that *challenges established model structures and data collection approaches* is required (cf. Refsgaard and Knudsen, 1996; Beven, 2001a,b; Young, 2001).
(2) More *tests of the internal-consistency of model predictions* (e.g. permeability distributions, water-table elevation, saturated area extent) would be helpful. Indeed, validation is greatly enhanced when several such data-series are available, as part of what has been described as multi-response validation (Mroczkowski *et al.*, 1997; Kokkonen and Jakeman, 2001).
(3) Hydrological processes and/or water *pathways that dominate at the hillslope scale* (so-called 'effective hillslope-scale processes') need to be identified more robustly (Bonell, 1998). Possible methodologies that may help are replicated, whole-hillslope tests (e.g. Lancaster, 2000) and models that explicitly address the source of rainfall–runoff non-linearity (Young, 2001).
(4) We need more assessments of whether complex physics-based approaches are better than *parsimonious, transfer-function approaches (e.g. DBM-model) based on 'good experience' (i.e. a range of case studies) at predicting the effects of land-use change* on rainfall–runoff processes.

Clearly, addressing the issue of how best to characterise those tropical soil parameters that regulate catchment rainfall–runoff behaviour is an important precursor to the reliable prediction of how land-use change might alter tropical soil parameters and thence streamflow generation and nutrient transport.

APPENDIX 31.1 GLOSSARY OF KEY MODELLING TERMS

Block permeability	An estimate of the soil or rock permeability typically derived by statistical manipulation of measured values to give estimates representative of a much larger scale (i.e. 'up-scaling'; see Wen and Gómez-Hernández, 1996).
Catchment parameters	Properties of a catchment that are largely unchanging with time (e.g. soil-rock permeability, porosity), but may be spatially variable.
DBM approach	An approach to modelling that incorporates transfer identification, with objective statistical evaluation and physical interpretation (see Young, 2001)
Effective parameters	Normally, values of parameters that give acceptable model simulations, and can be different to those from values measured in the field.
Grid-scale	In order to perform distributed (or semi-distributed) simulations with a physics-based model, the catchment is often divided into 100s or 1000s of elements in plan (Chappell and Ternan, 1992). The area of one of these elements is the grid-scale.
Impulse response function	The form of the output given a unit input (cf. Young, 1984).
Index of hydrological similarity	Catchment elements with the same index are assumed to have the same hydrological behaviour (cf. Beven, 2001a).
Model efficiency	How well the model output fits some statistical measure of the measured system output (i.e. the 'objective function'). Here the Nash and Sutcliffe (1970) efficiency measure is used (i.e. the one minus the ratio of the variance of the errors to the variance of the observed data, but then sometimes presented as a per cent).
Model parameterisation	The process by which parameters, such as soil parameters, are associated with each component of the model structure.
Model structure	The number of parameters in a model and their functional relationships with each other and the input-output variables.
Monte Carlo simulation	This is a process whereby 100s to 10 000s of simulations are undertaken with the same model structure, but with values of each parameter randomly selected from a normal or uniform distribution of values within an observed or 'realistic' range.
Natural soil pipes	Natural conduits or tunnels in the soil, formed by the action of water. Examples ranging in size from 0.05 m to 2 m diameter can be found within the humid tropics (Jones, 1990).
Parameter inversion	In the case of rainfall–runoff modelling, the identification of values of a particular parameter that is consistent with the model structure chosen and a good model efficiency.
Parameter uncertainty	Where values of a particular model parameter are derived by the 'parameter inversion' process, one set of parameters may give the same lumped output (e.g. riverflow) as a very different set, due to interaction between the model parameters. This problem and the subsequent uncertainty becomes worse as model complexity increases.
Pathline	The route taken by a fluid particle through the subsurface system.
Percolines	Zones of the soil, much broader than individual pores, natural soil pipes or fractures, where lateral water movement is much higher than that in the surrounding soil (after Bunting, 1961).
Permeability	The term is commonly used instead of the Coeficient of Permeability or more strictly the Saturated Hydraulic Conductivity. It is defined as the velocity of subsurface water through a unit area of saturated media under a unit hydraulic gradient (Darcy, 1856). It is distinct from (i) the term Unsaturated Permeability, which is the velocity of subsurface water through a unit area of unsaturated media under a unit hydraulic gradient, and (ii) the term Intrinsic Permeability, which is the intrinsic characteristic of the media, independent of the fluid (e.g., fresh water, salt water, oil) passing through it (Hubbert, 1940).
Preferential flow	The movement of one component of subsurface flow at a much greater velocity than the surrounding soil, and includes phenomena such as pipeflow, fissure flow and flows in percolines.

Sensitive model parameter	A model parameter, that if varied within the observed range of spatial variability would result in a large change in the model output (e.g., riverflow).
Transfer function	Basic *z*-domain representation of a linear digital filter between input(s) and output(s), expressing the filter as a ratio of two polynomials. See Middleton (2000) for an introduction to the topic of transfer function identification.
Up-scaling	Derivation of larger-scale estimates of a parameter (e.g., permeability) through rigorous statistical or numerical modelling of small-scale observations (see Wen and Gómez-Hernández, 1996).

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