TOPMODEL: A PERSONAL VIEW

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ABSTRACT

The minimum set of assumptions underlying TOPMODEL are explored, together with conditions under which they can or cannot be relaxed. It is concluded that it may be necessary to move towards spatially explicit solutions of the governing equations if the underlying q–D relationships are to be modified.

The simplicity of TOPMODEL invites its use as a submodel within a range of geomorphological and ecological models that are driven by hydrology. Some example applications are outlined for both soil erosion and solution.

INTRODUCTION

In the late 1960s, it became clear that the standard forecasting models used in engineering hydrology, typified by the Stanford Watershed Model (Crawford and Linsley, 1966), were too complex to be applied objectively, and that the processes that they represented were not always based on physical realities observed in the field, particularly in humid areas. The work on subsurface flow by Hewlett (Hewlett and Hibbert, 1963), Betson (1964), Whipkey (1965) and others accorded more with the observations of those involved in field studies, and referred to processes that were not included in existing hydrological models. For a geomorphologist, the existing hydrological models, by not representing real processes, also provided a particularly poor basis for understanding the critical role of surface and subsurface water in driving hillslope sediment transport.

This intellectual climate encouraged a search for novel hydrological models that were both simpler in structure, and had a closer relationship to the physical processes observed. Perhaps the critical concept was that of the partial contributing area, which was responsible for generating most of the flood peak runoff, and which came to be identified with the observed extent of surface saturation around the head of streams in humid areas. Part of the rationale for TOPMODEL was as a crucial product of that search.

In seeking simpler hydrological models, it was also recognized that all hydrographs seem to share a common form, of a rapid rise to peak and a slower decline, so that the ideal forecasting model should not be disproportionately complex. Strategies for retaining simplicity included the use of a minimum number of optimized parameters. The complement of this simplicity was to make best use of reliably measured data, and, for the catchment, the topography was the obvious candidate for full exploitation. Although, when TOPMODEL was first presented, topography had to be surveyed in the field, it was still by far the most accessible data set for the catchment, both in terms of measurement and of a clear visual understanding of its data structure. The semi-distributed form of TOPMODEL makes full use of topographic data, and it is only recently that we have begun to obtain field evidence that shows the strengths and limitations of what were, originally, a set of theoretical concepts.

In discussing TOPMODEL, this paper first discusses one of the possible rationales for the conceptual basis, and second how TOPMODEL may be able to realize its potential, not only within hydrology, but also as a hydrological driver for geomorphological process models for both sediment and solute transport.
THE RATIONALE OF TOPMODEL

The model could be approached from a number of theoretical directions. Here we start from the continuity equation for the plan area of a flow strip of variable width $w$, in which the horizontal distance, $x$, follows a curvilinear path down the line of greatest slope (see Figure 1). No assumptions about the nature of the flow are being made at this stage, and assumptions will be introduced only as and when strictly necessary.

\[
\text{inflow} - \text{outflow} = \text{net decrease in soil moisture deficit}, \quad D
\]

\[
dt[waj - (wa + d\tilde{a})(j + dj) + iw\ dx] = -wdD\ dx
\]

where $q = aj =$ discharge per unit contour width, $j =$ discharge per unit area = runoff rate and $i =$ net rainfall intensity.

Reversing the sign and dividing throughout by $w\ dx\ dt$,

\[
a \frac{\partial j}{\partial x} + \frac{j}{w} \frac{\partial (aw)}{\partial x} - \frac{\partial D}{\partial t} = i
\]

Using the geometric identity,

\[
\frac{1}{w} \frac{\partial (aw)}{\partial x} = w
\]

we then have:

\[
a \frac{\partial j}{\partial x} - \frac{\partial D}{\partial t} = (i - j)
\]

(1)

This equation is a completely general statement of hydrological continuity for the flow strip. The first two assumptions we choose to make are that we can express flow strip discharge as: (i) proportional to slope gradient; and (ii) some function of soil moisture deficit, $D$.

Soil moisture deficit is referenced to zero at soil saturation in what follows. The assumption that flow is proportional to water table gradient is equivalent to Darcy’s law, and is appropriate for laminar flow, and hence a good approximation for most subsurface flow except in large macropores. For overland flow, the assumption is less valid, and most versions of TOPMODEL separate the overland flow, at least partly for this reason. In TOPMODEL, it is also normally assumed that piezometric gradient is equal to (or perhaps in fixed proportion to) hillslope surface gradient. This convenient assumption breaks down significantly where
and/or when piezometric gradients are low. For such situations, the piezometric gradient itself varies
significantly with discharge, and the assumption made below, that local gradient is fixed over time, breaks
down.

The assumption that flow depends on a single valued function of soil water deficit is also an approximation,
which is strictly valid only if the moisture distribution in the soil is always the same for a given deficit. As a
first-order approximation, this may be adequate, but it is not difficult to devise significant exceptions. For
example, the same total deficit may be obtained after several days drainage; or after longer drainage followed
by near-surface wetting. These cases are usually distinguished in versions of TOPMODEL by using a separate
near-surface store, representing unsaturated water that contributes little to downslope flow.

These assumptions may be summarized in the equation:

\[ q = aj = \Lambda f(D) \]

for a suitable function \( f \). Formally, we may invert this to give:

\[ D = \Phi(aj/\Lambda) \]

for another function \( \Phi \), where \( \Lambda = \) slope gradient.

Next, we make the third, and most crucial assumption, that runoff (i.e. discharge per unit area) is spatially
uniform. This corresponds well with field observations in many small humid catchments, both around
hydrograph peaks and under low-flow conditions. This assumption requires that the \( \partial j/\partial x \) term in
Equation (1) is zero, since \( j \) does not change with position. For a consistent solution to the equation, the net
rainfall addition, \( i \), and the \( \partial D/\partial t \) term, must also be spatially invariant (i.e. must not change with position in
the catchment). Substituting for \( D \) in Equation (1) above:

\[ \frac{\partial D}{\partial t} = \frac{\partial \Phi(aj/\Lambda)}{\partial (aj/\Lambda)} \frac{\partial (aj/\Lambda)}{\partial t} \]

with

\[ \frac{\partial (aj/\Lambda)}{\partial t} = (aj/\Lambda) \frac{1}{j} \frac{\partial j}{\partial t} \]

if \( a, \Lambda \) do not vary over time at any point. We note that the term

\[ (aj/\Lambda) \frac{\partial \Phi(aj/\Lambda)}{\partial (aj/\Lambda)} \]

is itself a function of \( (aj/\Lambda) \) and that this term must also vary with time alone, and not with position. The
only suitable form for this term is, in general, a constant, which we will call \( -m \). The function \( \Phi \) must then
take the form of a logarithm. At its most general:

\[ D = \Phi(aj/\Lambda) = -m \ln(aj/\Lambda q0) = -m \ln(j) - m \ln(aj/\Lambda q0) \]

\[ \frac{\partial D}{\partial t} = \left[ (aj/\Lambda) \frac{d \Phi(aj/\Lambda)}{d (aj/\Lambda)} \right] \cdot \frac{1}{j} \frac{\partial j}{\partial t} = -m \frac{1}{j} \]

which is independent of position.

Substituting back into Equation (1), we obtain an alternative general form for the continuity equation:

\[ a \frac{\partial j}{\partial x} - \left[ (aj/\Lambda) \frac{d \Phi(aj/\Lambda)}{d (aj/\Lambda)} \right] \cdot \frac{1}{j} \frac{\partial j}{\partial t} = (i - j) \quad (1a) \]
This form can also be interpreted directly as a kinematic wave equation, with a celerity for the hydrograph of

\[ c = -\frac{aj}{(aj/L)d\Phi(aL/\Lambda)d(aL/\Lambda)} \]

For the logarithmic form, we then obtain the appropriate aspatial form of the continuity equation:

\[ \frac{dj}{dt} = \frac{j}{m}(i - j) \]  \hspace{1cm} (2)

together with the relationship between deficit and runoff,

\[ D = -m \ln(aj/\Lambda q_0) \quad \text{or} \quad q = aj = \Lambda q_0 \exp(-D/m) \]  \hspace{1cm} (3)

These define both the discharge \((q = aj)\) and the soil water deficit to saturation \((D)\) at every point in the catchment. With \(D\) chosen as zero at soil saturation, \(q_0\) is the discharge per unit width at saturation on unit slope gradient. This may vary from place to place within the catchment without violating the other assumptions.

We can also define the runoff required to produce local saturation (which varies from place to place). Setting \(D = 0\) in Equation (3) gives \(j_0 = \Lambda q_0/a\).

\(m\) is a soil depth parameter, invariant over the flow strip and over time, which shows how quickly discharge falls off with depth. At deficit \(D\), the lateral saturated hydraulic conductivity is

\[ K(D) = \frac{q_0}{m} \exp(-D/m) \]

The aspatial solution is only appropriate after any initial transients are eliminated. Their survival can be estimated from the wave celerity, which for this case is \(c = aj/m\).

Reasonable values for these terms (e.g. \(j = 1\) mm h\(^{-1}\); \(m = 10\) mm; \(a = 1000\) m; \(c = 100\) m h\(^{-1}\)) suggest that few initial transients would survive the first storm, except perhaps at locations very close to the divides. Once these transients have been eliminated, no more will be introduced provided that net rainfall \((i)\) remains spatially uniform. This condition is generally met most fully close to the divides, and tends to break down in the wetter hollows, where, conveniently, fresh transients survive most briefly. Near the divides, slope profiles are generally convex, so that flow depths are everywhere similar, giving spatially uniform infiltration delays over the areas. In the hollows, saturation deficits are decreasing rapidly, so that infiltration delays are also decreasing, with the result that net additions to the saturated zone, \(i\), are not spatially synchronous.

Equation (2) must be solved separately for zero and non-zero rainfall intensities. It is also helpful to solve it for the average runoff \(\bar{j}\) over a finite time interval \(\Delta t\), to allow computer iteration over reasonable time periods. The results are quoted here without derivation.

**Case 1:** \(i \neq 0\) (but constant over interval 0–\(t\))

\[ j = \frac{i}{1 + (i/j_s - 1) \exp(-i\Delta t/m)} \]  \hspace{1cm} (4)

\[ \bar{j} = m/\Delta t \ln[(1 - j_s/i) + j_s/i \exp(i\Delta t/m)] \]

**Case 2:** \(i = 0\)

\[ j = \frac{j_s}{1 + \Delta tj_s/m} \]  \hspace{1cm} (5)

\[ \bar{j} = m/\Delta t \ln(1 + \Delta tj_s/m) \]
where \( j_s \) is, in both cases, the runoff at the start of the time period. The storage at the end of the time period can be calculated directly from the instantaneous runoff, \( j \).

A final equation that can be useful is given by substituting the logarithmic relationship of Equation (3) directly into the generalized continuity Equation (1a). We then have:

\[
a \frac{\partial j}{\partial x} - \frac{m}{j} \frac{\partial j}{\partial t} = (i - j)
\]

In this expression, the assumption of spatially uniform runoff has been relaxed, and has been replaced by assuming the logarithmic relationship of Equation (3). In this form, neither runoff nor net rainfall need be spatially uniform, but assumptions (i) and (ii) above are still required.

In the discussion above, no account has been taken of the processes by which net rainfall is applied to the moisture deficit, \( D \). Most versions of TOPMODEL make use of unsaturated zone store, which delays the arrival of storm rainfall in the saturated deficit zone. The behaviour of this unsaturated store is clearly critical in determining whether the net rainfall addition term, \( i \), is assumed to remain spatially uniform. The present discussion does not consider the behaviour of the unsaturated store, except as a possible source of spatial non-uniformity of inputs.

**RELAXATION OF THE TOPMODEL ASSUMPTIONS**

Recently, there has been some discussion (e.g. Ambroise et al., 1996; and in this volume) about ways to relax the assumptions inherent in the simplest conceptual form of TOPMODEL. Some of these do not violate any of the assumptions above, while others propose changes that have generally focused on the form of Equation (3) above. In examining the effect of changing our assumptions, we need to examine the magnitude of any terms in Equation (1) that are being ignored. The duration of transient effects can also be followed through the kinematic wave celerity, \( c \).

Some relaxations of the simplest version of TOPMODEL do not violate principles or assumptions, although they may substantially increase the data requirements of the model. First, the saturated discharge \( (q_0) \) or near-surface hydraulic conductivity \( (K_0 = q_0/m) \) of the soil may be allowed to vary freely over the area of the catchment. The topographic index \( (a/\Lambda) \) for each point is then replaced by a soil–topographic index \( (q_0a/\Lambda) \), but no assumptions are violated. Where there are large local differences in \( q_0 \), however, there will be corresponding differences in deficit level, which may violate the assumption that the water table parallels the ground surface, and may lead to spatial difference in potential evapotranspiration. This may lead to local disconnection of the flow paths under dry conditions, with re-establishment of through linkage when the catchment wets up (e.g. Barling et al., 1994).

Secondly, the catchment may consist of a number of flow strips in parallel, each with different values for the soil depth parameter, \( m \). Although it may not be practicable to assign different values of \( m \) for each flow strip, it may be realistic to partition the catchment into two, or a few, distinct areas. For example, the Mendip East Twin catchment (Calver et al., 1972) was seen to consist of two roughly equal areas: one a poorly drained headwater area, and the other consisting of well-drained brown earth soils. This catchment generates a hydrograph with peaks from the two parts of the catchment that are commonly 24–48 hours apart, corresponding to \( m \) values that differ by a factor of 3–5. Most catchments are less extreme, but it may be worthwhile to distinguish \( m \) values for the stream head hollow and for the side slopes.

A third strategy is to allow parameter values to vary gradually over the year, maintaining water balances as values of \( m \) and/or \( q_0 \) vary. Provided that values change gradually, at rates that are slow compared with hydrograph response, the terms ignored are clearly negligible.

A fourth strategy, which has more serious implications for the assumptions, is to vary the form of the relationship between deficit and local discharge [Equation (3)]. The exponential equation of TOPMODEL has been replaced with a power law, usually of first or second order. The comparative properties of some alternative formulations are as shown in Table I. The forms in Table I have been normalized to give the same
saturated discharge, \( q_0 \), at unit gradient, and the same saturated conductivity at the surface, \( q_0/m \). It may be seen that the form of the parabolic case converges on the exponential case as the exponent, \( r \), tends to infinity. The parabolic case has normally been used for the case of \( r = 2 \).

It may be seen that the kinematic wave celerity behaves in a very different way for the linear and parabolic cases. Whereas for TOPMODEL, transients only survive near divides, where spatial differences in input are least, the linear and parabolic cases give slowest wave velocity, and therefore longest duration of transients, where gradient is low along the valley axis. The spatial differences in input in this zone may therefore have a significant effect on the output hydrograph, suggesting that the spatial terms in Equation (1) cannot properly be ignored. It is not, of course, difficult to solve the full equation without assuming spatially uniform runoff, but it has to be solved for a particular topographic setting. In this form it still retains many of the advantages of TOPMODEL in having a low number of parameters that require calibration or optimization, and an interpretation in terms of moisture deficits throughout the catchment.

In some cases, the relationship between \( q \) and \( D \) during the recession curve has been used to discriminate between models. Computer simulations of recession curves, with zero inputs, show that full solutions with TOPMODEL assumptions can emulate closely the quadratic \( q - D \) relationship associated with the parabolic model as a result of transients that survive from the start of the recession period, as illustrated in Figure 2. Similarly, simulations with the parabolic form show the increasing importance of transients that are generated within the recession period, providing much more misleading \( q - D \) relationships than those illustrated in Figure 2. Another source of error may arise where recession curves are generated for periods with non-zero evapotranspiration.

**TOPMODEL IN THE CONTEXT OF A SOIL BULK DENSITY MODEL**

Although TOPMODEL provides a satisfactory conceptualization for the saturated zone, it is less effective for the unsaturated zone. One novel way to begin to resolve this difficulty is by considering the bulk density profile of the soil explicitly. It is argued here that the bulk density profile in a natural soil is the result of a balance between upward diffusive movements and a downward settlement under gravity. This balance is represented by the expression:

\[
\Omega(z, e) = -\kappa \frac{dz}{de}
\]

**Table I. Comparison of properties of some alternative formulations of TOPMODEL**

<table>
<thead>
<tr>
<th></th>
<th>Exponential (TOPMODEL)</th>
<th>Linear</th>
<th>Parabolic</th>
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<tbody>
<tr>
<td>Discharge, ( q = aj )</td>
<td>( q = q_0 \Lambda \exp(-D/m) )</td>
<td>( q = q_0 \Lambda (1 - D/m)^r )</td>
<td>( q = q_0 \Lambda \left(1 - \frac{D}{mr}\right)^\uparrow )</td>
</tr>
<tr>
<td>Deficit = ( \Phi(aj/\Lambda) )</td>
<td>( D = -m \log(aj/\Lambda) )</td>
<td>( D = m[1 - (aj/\Lambda q_0)]^r )</td>
<td>( D = mr \left[1 - \left(\frac{aj}{q_0 \Lambda}\right)^{1/r}\right] )</td>
</tr>
<tr>
<td>Saturated conductivity</td>
<td>( K = \frac{q_0}{m} \exp(-D/m) )</td>
<td>( K = \frac{q_0}{m} )</td>
<td>( K = \frac{q_0}{m} \left[1 - \left(\frac{D}{mr}\right)^{r-1}\right] )</td>
</tr>
<tr>
<td>( -(aj/\Lambda) \frac{d\Phi(aj/\Lambda)}{d(aj/\Lambda)} )</td>
<td>( m )</td>
<td>( m \left(\frac{aj}{q_0 \Lambda}\right)^{1/r} )</td>
<td>( \frac{1}{m} (aj)^{1-1/r}(q_0 \Lambda)^{1/r} )</td>
</tr>
<tr>
<td>Wave celerity, ( c )</td>
<td>( aj/m )</td>
<td>( q_0 \Lambda/m )</td>
<td>( \frac{1}{m} (aj)^{1-1/r}(q_0 \Lambda)^{1/r} )</td>
</tr>
</tbody>
</table>

\* Valid for \( 0 < D < m \).
\^ Valid for \( 0 < D < 2m \).
where $\Omega$ is the rate of settlement, $\varepsilon$ is the effective available porosity, $\kappa$ is the mixing diffusivity and $z$ is depth into the soil. The form of $\Omega$ is not well known. It might be thought to depend on both overburden pressure, 

$$
\int_0^z (1 - \varepsilon) \, dz'
$$

and/or on porosity directly. The simple form that appears to be most compatible with TOPMODEL is $\Omega = \gamma \varepsilon^2$. This leads to the following solutions in the context of TOPMODEL:

$$
\begin{align*}
\varepsilon &= \varepsilon_0 / (1 + z/z_0) \\
D &= m_0 \ln(1 + z/z_0) \\
q &= q_0 \Lambda \exp(-D/m) = q_0 \Lambda (1 + z/z_0)^{-m_0/m} \\
K_{\text{SAT}} &= \frac{\varepsilon_0 q_0}{m} \left( 1 + \frac{z}{z_0} \right)^{-(1+m_0/m)}
\end{align*}
$$

where $m_0 = \kappa / \gamma = \varepsilon_0 z_0$.

It may be reasonable to assume that $m_0 = z_0 = m$ and that $\varepsilon_0 = 1$, in which case the expressions become somewhat simpler. The diffusive mixing is also partly responsible for soil creep, which transports soil through the profile at a total integrated transport rate of $\kappa \varepsilon_0 \Lambda$. The choice of form for the settlement rate has implications for the integration with TOPMODEL. For example, a linear settlement law, $\Omega = \gamma \varepsilon$, leads to an exponential decline in porosity with depth into the soil, and a maximum total possible deficit of $\varepsilon_0 m_0$. This form is more compatible with the linear or parabolic modifications of TOPMODEL, in which deficit has a similar upper bound.

The proposed form for the porosity profile allows the soil water deficit to build up, in principle to arbitrarily large values, although truncation at depth makes only very small changes to the forecast subsurface flows. In addition, a full representation of flow in the unsaturated zone requires a specification of the available pore size distribution associated with each effective porosity, $\varepsilon$, and this approach is not pursued further here. It is well known that soil bulk density profiles show an increase with depth which is broadly in accord with the profiles described here, although most data sets do not discriminate well between the alternatives proposed here.
TOPMODEL AS A SEDIMENT TRANSPORT MODEL

Short-term sediment transport models are commonly driven by tractive stress or flow power, which may be derived from the overall flow hydrology. As an illustration of the type of relationship used, sediment transport may be estimated (Kirkby, 1993) from the relationship:

\[ C \propto q_o(q_o \Lambda - \Theta) \]  

(8)

where \( C \) is the sediment transport capacity, \( q_o \) is the overland flow discharge per unit width and \( \Theta \) is a flow power threshold related to grain size, etc.

Other important factors may be included in such an expression, including a more detailed consideration of grain size and the recognition that not all sediment transport is at full capacity, but the simple form of Equation (8) is sufficient to illustrate the relevance of linkages with TOPMODEL. Perhaps the two most important types of linkage are the use of TOPMODEL to describe the distribution of overland flow over the hillslope, and its use to obtain the frequency distribution of overland flows over a period, and responses to possible global climate or land use change.

Because soil erosion tends to be most severe in semi-arid and agricultural areas, where Hortonian overland flow is important, most simple erosion models assume that overland flow is generated by infiltration excess, and builds up linearly downslope. As recent work has shown the importance of subsurface flow and saturation excess, at least seasonally, in many semi-arid environments, and as soil erosion is more widely recognized as a significant problem in many temperature areas, the relevance of forecasting subsurface flow and saturated conditions has increased. Clearly TOPMODEL forecasts a concentration of saturation overland flow near the base of hillslopes, particularly where they are concave in plan and/or profile. The sediment transport generated by selective soil erosion in these areas accentuates their form with a strong positive feedback, giving rise to pronounced plan-convergent hollows and slope profiles that are predominantly convex, with short, sharp basal concavities.

Where the soil surface is associated with a root mean square roughness, which may be associated with a fractal spatial dependence, TOPMODEL is also able to help define the depressions along which overland flow selectively reaches the surface. One approach that may be used here is to obtain the effective deficit by an iterative process as follows.

1. Calculate deficit and discharge relative to the mean surface level.
2. Recalculate the modified saturation level so that the sum of the overland flow (in the base of the depressions where the modified level intersects) and the subsurface flow (corresponding to the modified level) are equal to the previously calculated discharge. This calculation is only meaningful where overland flow velocity is greater than subsurface velocity.

For long-term slope evolution, this response of the flow and sediment transport to roughness also has to take into account the evolution of the RMS roughness over time.

For a given topographic location, defined by \( a/A \), the simplicity of TOPMODEL makes it possible to simulate readily the frequency distribution for the various components of the hillslope hydrology, many of which are relevant to sediment and solute transport rates. By assuming stochastic distributions of rainfall and potential evapotranspiration (or temperature), we may generate frequency distributions (Figure 3) calculated for periods of hundreds to thousands of years, for, among others:

- overland flow runoff (for soil erosion);
- subsurface runoff (for solute transport);
- actual evapotranspiration, vegetation biomass and organic soil (for soil resistance); and
- saturation deficit (for piezometric pressures in soil stability).
This approach provides a topographically sensitive way of estimating not only storm responses, but also long-term average responses and magnitude–frequency distributions for the rates of geomorphological processes. It also allows some estimates to be made about the possible responses to changes in climate or land use practice. Alternative approaches, which take a somewhat simpler view, include Kirkby (1978) and Ijjasz-Vasquez et al. (1992).

TOPMODEL AS A SOLUTE MODEL

Just as overland flow is mainly responsible for driving soil erosion, subsurface flow is the main driver for solute removal. Two important issues that TOPMODEL is able to address are those of residence times in the soil, and of the locus of solute equilibration within the soil profile.

A simplified view of the subsurface water column is that it shows no vertical mixing with water added from above. This corresponds well to observations of ‘plug flow’ in soils without well-developed macropores. Outflow is considered to take place by squeezing (vertically) and stretching (horizontally) the saturated layer. Addition of new storm water is by adding it as an even layer to reduce the existing deficit at every point. These assumptions ignore bypassing flow and diffusive mixture of the water layers, and the magnitude of these effects should be determined in particular cases.

Figure 4 shows an example of the way in which storm water, here added in a single instantaneous burst, is combined with pre-existing (‘old’) water in the saturated zone. It may be seen that even during the flood peak, defined as the period (2-2 days) required for total outflow to equal storm input, over 40% of the outflow is of old water, and that this fraction continues to rise. This means that at the end of the flood, the remaining 40% of the storm rain water is already 2-2 days old. Storm dilution effects are therefore short-lived in even this idealized case. In practice the residence time of flood outflows is also increased both by any diffusive mixing and the time to infiltrate through the unsaturated zone. It is argued that the combined effect of these processes is to ensure that even flood flows represent residence times long enough to attain equilibration of most soil solutes with soil water, requiring perhaps 100 hours. It should be noted that the residence times suggested here are very much shorter than those implied in Wolock et al. (1989).

By defining the level of flow in the soil, TOPMODEL also helps to simulate the level at which equilibration occurs. Since the regolith is, in general, progressively less weathered with depth as it approaches intact

bedrock, the equilibrium concentration of solutes in the water tends to increase with depth. At depth, the upward diffusion of the solutes is more rapid than the rate at which they are advected away by the flow (Kirkby, 1985). Closer to the surface, advection dominates and the solution rate is the product of lateral flow rate (increasing towards the surface) and equilibrium concentration (increasing with depth). TOPMODEL provides a means of understanding how solution rates vary over the hillside with these factors. It also provides forecasts of the total amount of water that is available for solution (subsurface flow) and for soil erosion (overland flow), showing, for example, that solution rates are likely to fall and soil erosion rates to rise in saturated headwater hollows.

CONCLUSION

It is important to stretch the concepts that lie behind TOPMODEL, and explore whether the strengths of the underlying approach can be retained in a more general framework. At present, there is a need to explore more fully the magnitude of transient terms that are being ignored, and to decide whether to move from the aspatial simplicity of TOPMODEL to a spatially distributed solution of Equation (6) or a version of it from Table I, which, while requiring spatially explicit solutions, still retain the low number of parameters that give TOPMODEL part of its attractiveness in comparison with previous hydrological models.

The geomorphological applications for TOPMODEL have been described here only in outline. In limited space, it has not been the intention to pursue them in detail, but to demonstrate one of the great values of the TOPMODEL approach: that its simplicity and physical basis invite incorporation into a wide range of geomorphological and ecological models in which hydrology is an important driving factor.

REFERENCES


