The versatile integrator of surface atmospheric processes
Part 2: evaluation of three topography-based runoff schemes

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Abstract

Three different schemes of topography-based runoff production [versatile integrator of surface atmospheric processes (VISA)-TOP1, VISA-TOP2, and VISA-TOP3] are described for a land-surface model (LSM) developed for use with a general circulation model (GCM). The schemes’ sensitivities to some key parameters are assessed for two catchments using data sets developed for the Project for Intercomparison of Land-Surface Parameterization Schemes (PILPS) Phase 2e. VISA-TOP1 differs from VISA-TOP2 only in how to treat oversaturated soil water from the soil layers. In VISA-TOP1, the oversaturated soil water is thrown out of the soil column; hence, it no longer plays a role in the ensuing soil water budgets. In VISA-TOP2, this oversaturated soil water is recharged back to the unsaturated soil layers above the water table; hence, it continues to involve in the water budgets. Unlike VISA-TOP1 and VISA-TOP2, VISA-TOP3 relaxes its dependence on the topographic parameters. The oversaturated soil water is treated the same in both VISA-TOP2 and VISA-TOP3. All three models reproduce the daily and seasonal cycles of streamflow provided that different values of the saturated hydraulic conductivity decay factor are used. The decay factor controls the timing and partitioning of subsurface runoff. In both VISA-TOP1 and VISA-TOP2, an anisotropic parameter explaining different hydraulic conductivities in the vertical and horizontal directions is critical for using the topographic index in the land-surface model. In the VISA-TOP2 scheme, the topography-controlled subsurface runoff is dominant because the oversaturated water is recharged to upper unsaturated soil layers to raise the water table. The water budgets in all these schemes show dramatically different responses to the decay factor, indicating that the calibrated parameters and the model formulations should not be separated.

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1. Introduction

Parameterization of runoff-related processes in soil vegetation–atmosphere transfer (SVAT) schemes has been an active research topic. There is an increasing awareness that the parameterization needs to explicitly account for the topographic control on the soil moisture distribution and the runoff production.

Beven and Kirkby (1979) were among the first to develop a simple conceptual model at the hillslope and the catchment scales to model the runoff production using information on hillslope topography and...
soil and rainfall variabilities. The so-called TOPography-based hydrological MODEL (TOPMODEL) formalism has undergone significant enhancements over the years (Beven and Kirkby, 1979; Beven et al., 1994, 1995). Recently, the basic TOPMODEL approach has been employed by various SVAT schemes developed for use with general circulation models (GCMs) (Famiglietti and Wood, 1991; Stieglitz et al., 1997; Koster et al., 2000; Dulachet al., 2000; Chen and Kumar, 2001). Yang et al. (2000) implemented a topography-based runoff scheme into the common land model (CLM) (Dai et al., 2003), and the resulting model was tested with the meteorological and streamflow data from the Red-Arkansas River basins.

It is well recognized that TOPMODEL involves a considerable amount of uncertainties due to its basic assumptions, meaning of the model parameters, and derivation of topographic index distributions from digital terrain data at different resolutions (Beven, 1997). However, it is still attractive for climate modelers to use the TOPMODEL concept to parameterize the topographic effects in SVATs for climate studies. Such parameterization may involve three steps: (1) implementing the TOPMODEL concept in a SVAT model, testing alternative formulations, and determining sensitive parameters in offline mode at the hillslope and catchment scales; (2) testing model formulations and determining parameters in offline mode on a global basis; and (3) testing model formulations, determining parameters, and studying the surface water–climate interaction with GCMs. This paper addresses the first step.

Yang and Niu (2003-this issue), hereafter referred to as Paper 1, describes an integration of recent advances in snow physics, topography-related runoff processes, and leaf growth into the National Center for Atmospheric Research (NCAR) Land-Surface Model (LSM) of Bonan (1996). The resulting model is called the versatile integrator of the surface atmospheric (VISA) processes. In the present paper, we describe three parameterization schemes of topography-related runoff in VISA. The schemes’ sensitivities to some key parameters are assessed for two catchments using data sets developed for the Project for Intercomparison of Land-Surface Parameterization Schemes (PILPS) Phase 2e (Bowling et al., 2003-this issue).

### 2. Parameterization of topographic effects on runoff production: VISA-TOP1

While Paper 1 has detailed a topography-related parameterization scheme of runoff in VISA, this section describes a variant of the runoff scheme, hereafter referred to as VISA-TOP1. In order to focus on the topographic effects, the VISA model in this paper does not include the leaf growth component described in Paper 1.

#### 2.1. Saturated hydraulic conductivity

The saturated hydraulic conductivity decreases exponentially with depth according to

\[ K_{sat}(z) = K_{sat}(0) e^{-\frac{z}{f}} \]  

where \( K_{sat}(0) \) is the surface value of saturated hydraulic conductivity, and \( 1/f \), the e-folding depth of \( K_{sat} \), can be determined through sensitivity analysis or calibration against the recession curve of the observed streamflow. Table 1 lists values of \( f \) used in the literature.

#### 2.2. Surface runoff

Surface runoff consists of overland flow by the Dunne mechanism which requires rainfall to impinge on a saturated ground surface and overland flow by the Horton mechanism which is generated when rainfall rate exceeds the infiltration capacity of the soil. The mathematical representation of the above processes takes the form of

\[ R_s = F_{sat} Q_{wat} + (1 - F_{sat}) \max(0, (Q_{wat} - I_{max})) \]  

where \( Q_{wat} \) is the recharging rate at the soil surface, \( I_{max} \) is the soil infiltration capacity dependent on soil

<table>
<thead>
<tr>
<th>Authors</th>
<th>Sites/regions</th>
<th>( f )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Beven, 1982</td>
<td>Various sites</td>
<td>–2.35–9.15</td>
</tr>
<tr>
<td>Famiglietti et al., 1992</td>
<td>FIFE</td>
<td>1.5–5.17</td>
</tr>
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<td>Stieglitz et al., 1997</td>
<td>Sleepers River</td>
<td>3.26</td>
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<tr>
<td>Dai et al., 2002</td>
<td>Global</td>
<td>2.0</td>
</tr>
<tr>
<td>Chen and Kumar, 2001</td>
<td>North America</td>
<td>1.8</td>
</tr>
<tr>
<td>Yang et al., 2000</td>
<td>Red Arkansas River</td>
<td>8.0</td>
</tr>
</tbody>
</table>
texture and moisture conditions (Entekhabi and Eagleson, 1989), and \( F_{\text{sat}} \) is the saturated fraction, which is determined by the topographic characteristics and soil moisture state of a grid cell,

\[
F_{\text{sat}} = \frac{1}{f_{z_j}} \int_{\lambda \geq (\bar{\lambda} + f_{z_j})} \text{pdf}(\lambda) d\lambda
\]  

(3)

where \( \lambda = \ln(a/\tan \beta) \) is the topographic index where \( a \) is the contribution area and \( \tan \beta \) is the local slope; \( \bar{\lambda} \) is the mean value of \( \lambda \) in the grid cell; \( \text{pdf}(\lambda) \) is the probability density function of \( \lambda \); and \( z_j \) is the grid-mean water table depth.

2.3. Subsurface runoff

Subsurface runoff is parameterized as,

\[
R_{sb} = R_{sb,\text{TOP}} + R_{sb,\text{BOT}} + R_{sb,\text{SAT}}
\]

(4)

where \( R_{sb,\text{TOP}}, R_{sb,\text{BOT}}, \) and \( R_{sb,\text{SAT}} \) represent productions of subsurface runoff due to topographic control, bottom drainage, and saturation excess, respectively.

2.3.1. Topographic control

Following Sivapalan et al. (1987), the production of subsurface runoff due to topographic control is

\[
R_{sb,\text{TOP}} = \frac{a}{f_{z_j}} K_{\text{sat}}(0) e^{-f_{z_j}}
\]

(5)

Fig. 1. (a) The topographic index computed from 1000 m DEMs and adjusted according to Eq. (8) for Øvre Abiskojokk (seven grid cells). (b) The topographic index computed from 1000 m DEMs and adjusted according to Eq. (8) for Øvre Lansjärvi (10 grid cells).
where \( a \) is an anisotropic factor accounting for the differences in the saturated hydraulic conductivities in the lateral and vertical directions introduced by Chen and Kumar (2001) to simulate the desired streamflow response.

### 2.3.2. Bottom drainage

The gravitational loss of soil water from the bottom of the model soil column is given by

\[
R_{\text{sb,BOT}} = k_6 + \left( \theta_6^{N+1} - \theta_6^N \right) \left( \frac{\partial k}{\partial \theta} \right)_6
\]

where \( k_6 \) is the hydraulic conductivity at the bottom of the sixth (i.e., the lowest) soil layer, and \( \theta_6^N \) is the volumetric soil moisture within the sixth soil layer at time step, \( N \).

### 2.3.3. Saturation excess

The production of subsurface runoff due to saturation excess is

\[
R_{\text{sb,SAT}} = \sum_{i=1}^{6} \max[0, ((\theta_i - \theta_{\text{sat}}) \Delta z_i / \Delta t)]
\]

where \( \theta_i \) and \( \theta_{\text{sat}} \) are the volumetric soil moisture of the \( i \)th layer and the soil porosity, respectively; and \( \Delta z_i \) and \( \Delta t \) are the soil thickness of the \( i \)th layer and the timestep, respectively.

### 2.4. Water table depth

The water table depth, \( z_f \), is computed following Chen and Kumar (2001). See Paper 1 for details.

### 3. The topographic index: calculation, distribution and fitting

The topographic index, \( \lambda \), is often computed from regularly spaced grids of elevation values called digital elevation models (DEMs). While the U.S. Geological Survey (USGS) provides 30-m resolution DEMs for many parts of the USA and about 100 m DEMs for the entire USA, only 1000 m DEMs are available for the entire Earth. It is widely recognized that the distribution of the topographic index is strongly dependent upon the DEM resolution. Wolock and McCabe (2000) proposed a regression equation to relate at 1000-m resolution to that at 100-m resolution as follows:

\[
\lambda_{100 \text{ m}} = 0.961 \times \lambda_{1000 \text{ m}} - 1.957
\]

We computed the topographic index in the Swedish Torne–Kalix basin (Bowling et al., 2003-this issue) based on the 1000 m DEM. The values were adjusted according to Eq. (8), and the results for two subbasins, the 566 km² Övre Abiskojokk and the 1341 km² Övre Lansjärv, are shown in Fig. 1. High values of topographic index are seen near the river channels and lakes (e.g., Grid Cell 4 of Övre Abiskojokk). The topographic parameters, i.e., mean, minimum, maximum, variance, and skewness of the topographic index in each grid cell, are listed in Table 1 for Övre Lansjärv and in Table 2 for Övre Abiskojokk. Following Sivapalan et al. (1987), a three-parameter gamma distribution function was used to fit the actual discrete distribution of the topographic index in each grid cell.

Fig. 2 compares cumulative distribution function (CDF) derived from the analytical three-parameter gamma distribution and from the actual discrete distribution. Basically, the gamma distribution captures the actual distribution for each grid cell especially when \( \lambda \) is greater than \( \lambda \). Actually, for the calculation of the saturated fraction using Eq. (3), it is not necessary to capture the distribution when \( \lambda \) is less than \( \lambda \). The maximum saturated fraction (when the mean water table depth is zero) is only determined by the topographic indices. For instance, \( \lambda \) for Grid 7 in Övre Lansjärv (9.84) corresponds to the CDF value of 38% which is

<table>
<thead>
<tr>
<th>Grid no.</th>
<th>( f )</th>
<th>Mean</th>
<th>Min.</th>
<th>Max.</th>
<th>Variance</th>
<th>Skewness</th>
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<td>3.64</td>
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<td>10</td>
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<td>6.70</td>
<td>19.64</td>
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<td>40.16</td>
</tr>
<tr>
<td>Average</td>
<td>3.93</td>
<td>10.49</td>
<td>6.95</td>
<td>19.01</td>
<td>6.68</td>
<td>39.43</td>
</tr>
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</table>
the maximum saturated fraction when \( z_T \) is zero. As \( z_T \) increases, the cumulative area for the topographic index greater than \( \bar{z} + f z_T \) decreases; hence, the saturated fraction is less than its maximum value. The analytical three-parameter gamma distribution is used in the model to improve the computational efficiency.

4. Model demonstration

4.1. Meteorological variables and land-surface characteristics

The data set used in this study is a long-term (1979–1998) hydrometeorological data collected over
the Torne–Kalix River system in Northern Scandina-
via, which has a combined area of 58,000 km² repre-
sented by 218 0.25° computational grid cells at hourly
time step. The meteorological forcing data and the
land-surface characteristics were described in detail in
Bowling et al. (2002). Nine vegetation types provided
in the PILPS 2e experimental instructions were com-
bined to form six land-surface types, each consisting of
up to four tiles per grid cell (Table 4). Out of the 218
grids in the entire Torne–Kalix river basin, there are
33 grid cells of surface type 1 which was mainly
distributed in the mountainous regions, 150 grid cells
of surface type 2 in the plain regions, and 31 grids of
surface type 3 in the mountain-plain transitional
regions. The fractional coverage of vegetation types
in each grid cell was taken from the PILPS 2e
instructions. The values of other vegetation parame-
ters, such as vegetation height, displacement height,
roughness length, architectural resistance, root distri-
bution, and monthly leaf area index, were specified
following the PILPS 2e instructions. The minimum
stomatal resistance, \( r\text{\_min} \), given in the instructions was
converted to the maximum stomatal conductance,
\( V\text{\_max} \), using a formula:
\[
V\text{\_max} = \frac{3 \rho_a}{c_f r\text{\_min}}
\]
(Dickinson et al., 1998), where \( \rho_a \) is air density and \( c_f \) is a
conversion factor \((7.46 \times 10^{-4} \text{ g mol})\). Thus, a value
of 120.0 m s\(^{-1}\) for \( r\text{\_min} \)is approximately equal to
39.3 \( \text{mol m}^{-2} \text{ s}^{-1} \) for \( V\text{\_max} \).
The soil thermal and hydraulic parameters, such as
heat capacity, thermal conductivity, hydraulic conduc-
tivity, and matrix potential at saturation, \( b \) parameter
(Clapp and Hornberger, 1978), and porosity, were cal-
culated from percent sand and percent clay using the
The statistics of the topographic index are given in
Tables 2 and 3. Because the saturated hydraulic
conductivity decay factor, \( f \), is a calibration parameter,
we seek a simple formula relating \( f \) calibrated against
the observed streamflow for two small catchments to
the entire Torne–Kalix basin through use of elevation,

\[
f = \text{max}(0.1, a - b z\text{\_top})
\]
where \( a \) and \( b \) are two adjustable coefficients, here,
\( a = 5.0 \), \( b = 0.004 \); and \( z\text{\_top} \) is the grid cell mean ele-
vation in meter. The catchment-averaged decay factor
is 3.93 for Övre Lansjär and 1.36 for Övre Abisko-
jokk, suggesting that the soil in the lowly lying Övre
Lansjärv catchment is more densely compacted than
that in the mountainous Övre Abiskojokk catchment.

### 4.2. Numerical results

The model was integrated for 11 years from 1988
to1998, and the last 10 years from 1989 to1998 were
analyzed for two catchments. Because the catchments
span multiple 0.25° grid cells, the modeled runoff
from the contributing grid cells was summed and
scaled according to the contributing area of each cell.
Fig. 3 compares the 10-year time series of the mod-
elled runoff with the observed streamflow. With the
parameters specified in Tables 2–4, VISA-TOP1
produces good simulations of streamflow. The surface
runoff is dominant in the beginning of the snow-
melting season, while the subsurface runoff is domi-
nant during the recession period.

During the melting or raining period, the soil
becomes wet, the water table rises, and the saturated
fraction is large (Fig. 4); hence, surface runoff is large.
Although Övre Lansjär has a shallower water table,
its saturated fraction is lower than that in Övre
Abiskojokk because the former has a larger \( f \) and a
smaller fraction of lakes. The saturated fraction in
Övre Abiskojokk varies from 20% to 32%, compared
to that from 10% to 22% in Övre Lansjär.

Not only is the hydrograph different in both catch-
ments, the proportions of subsurface runoff also show
distinct patterns. In Övre Lansjär, \( R_{\text{sb,BOT}} \) is zero,
whereas it is most pronounced in Övre Abiskojokk.
\( R_{\text{sb,SAT}} \) is dominant in Övre Lansjär, while it is
relatively modest in Övre Abiskojokk. \( R_{\text{sb,TOP}} \) is weak
in both catchments, especially in the lower Övre
Lansjär.
Fig. 3. (a) Time series of the modeled unrouted streamflow compared with the observed streamflow for Övre Abiskojokk. (b) Time series of the modeled unrouted streamflow compared with the observed streamflow for Övre Lansjärv.
The timing of peaks is different for $R_{sb,TOP}$, $R_{sb,BOT}$, and $R_{sb,SAT}$ as shown in Fig. 4. The topographic control peak has the same timing as the water table peak, which is similar to the timing of the saturation excess peak. The bottom drainage peak is much delayed because of a large time scale involved in the transport of soil water from the model surface to the bottom.

5. Sensitivity tests

Due to the uncertainties in determining $f$ and $z$, we will test the model’s sensitivities to these two parameters. In addition, the model’s sensitivities to the lake fraction and the topographic index are also assessed.

![Fig. 4](image-url)  
Fig. 4. Time series of the modeled unrouted streamflow, water table depth, saturated fraction, soil moisture at different layers, and subsurface runoff due to saturation excess, topographic control, and bottom drainage for Övre Abiskojokk (left panel) and Övre Lansjärv (right panel). The observed streamflow is also shown.
5.1. Sensitivity to saturated hydraulic conductivity decay factor, $f$

The decay factor, $f$, is the most important parameter in the TOPMODEL framework, because it is used in determining the vertical profile of saturated hydraulic conductivity and, hence, the vertical distribution of soil moisture, the saturated fraction, and the baseflow. It varies widely from $-2.35$ to $9.15$ in the literature (cf. Table 1). Its precise determination requires the observed vertical distribution of the saturated hydraulic conductivity (Famiglietti et al., 1992). It is hard to measure its horizontal distribution within a catchment or a grid cell in GCMs due to extremely heterogeneous soil textures. We have performed sensitivity tests here to investigate how $f$ affects the simulated recession curves and the partitioning of subsurface runoff.

The decay factor, $f$, is crucial for controlling the subsurface runoff timing. Fig. 5 shows the modeled unrouted subsurface runoff for various values of $f$ compared to the observed streamflow in Övre Länssjärv in 1992, a year characterized by strong snowmelt in spring and rain in fall. In the case of $f=2$, subsurface runoff suffers a severe time lag; for $f=3$, subsurface runoff is largely improved. The best fit occurs when $f=4$. The model is relatively insensitive to $f$ if it is greater than 4.

The decay factor affects the calculation of the water table depth and the saturated fraction. A larger $f$ leads to a faster decline of the saturated hydraulic conductivity with depth and, hence, to a smaller bottom drainage, which promotes a rise in the water table. However, the impact of increasing $f$ on the saturated fraction is not straightforward. An increase in the saturated fraction is associated with an increase

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**Fig. 5.** Time series of the modeled unrouted subsurface runoff, water table depth, and saturated fraction for different values of $f$ in Övre Länssjärv.
in $f$ from 2 to 3, whereas the saturated fraction starts to decrease if $f$ increases further.

The decay factor, $f$, is crucial in estimating $R_{sb, TOP}$. The 10-year averaged results of $R_{sb, TOP}$, $R_{sb, BOT}$, and $R_{sb, SAT}$ for various values of $f$ are shown in Fig. 6. $R_{sb, TOP}$ increases as $f$ increases from 2 to 8, whereas $R_{sb, SAT}$ decreases for the same range of $f$. $R_{sb, BOT}$ dominates if $f$ is less than 2. For the same $f$ value, $R_{sb, TOP}$ is always larger in Övre Abiskojokk than in Övre Lansjärvi because the former has a smaller mean topographic index. Although $f$ affects both surface and subsurface runoff components, total runoff is relatively not affected (Fig. 7). Notice that the Torne–Kalix basin lies above the Arctic Circle and, hence, is energy limited. How $f$ influences total runoff in midlatitudes and in tropics will be subject to future study.

5.2. Sensitivity to anisotropic factor

The anisotropic factor was first introduced by Chen and Kumar (2001) to account for the differences in the saturated hydraulic conductivities in the lateral and the vertical directions. They used $\alpha = 2000$, meaning that the saturated hydraulic conductivity is 2000 times as large in the lateral direction as in the vertical direction. Our test showed that the soil dries out quickly if $\alpha = 2000$ is used and that $R_{sb, TOP}$ is negligible if $\alpha = 1$ is used. In the control run, $\alpha = e^{2.5} = 12.2$ was used to fit the observed recession curve.

We have done five sensitivity experiments with $\alpha = e^0$, $e^1$, $e^2$, $e^3$, and $e^4$, respectively. As $\alpha$ increases from $e^0$ to $e^4$, $R_{sb, TOP}$ increases from 1.37% to 46.65% in Övre Lansjärvi and from 2.32% to 62.91% in Abiskojokk. However, subsurface runoff, surface run-
off, and total runoff change little (Table 5). In our model, \( z \) cannot exceed \( e^5 \) despite \( x = 2000 \) in Chen and Kumar (2001); otherwise, the soil would dry out to below zero.

### 5.3. Sensitivity to lake fraction

In the ‘without lake’ experiment, it is assumed that the saturated fraction for the lake tile is zero; hence, runoff is zero as in the original NCAR LSM (Bonan, 1996). In addition, the water table depth for the lake tile is prescribed at the model bottom (6.3 m). Fig. 8 shows the saturated fraction and the water table depth simulated with and without lake. Both variables show systematic reduction in the ‘without lake’ experiment and total surface runoff decreases by 3% and 28% in Övre Lansjärvä and Övre Abiskojokk, respectively.

### 5.4. Sensitivity to DEM resolution

Higher-resolution DEMs describes the contribution area and the local slopes more accurately. Because the topographic index computed from the 1000 m DEM is larger than that from the 100 m DEM by about 2 (cf.

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**Table 5**

Sensitivities to the anisotropic factor

<table>
<thead>
<tr>
<th>Subbasin</th>
<th>( z )</th>
<th>( R ) (m(^3)/s)</th>
<th>( R_s ) (m(^3)/s)</th>
<th>( R_{sb} ) (m(^3)/s)</th>
<th>( R_{sb,BOT} ) (%)</th>
<th>( R_{sb,TOP} ) (%)</th>
<th>( R_{sb,SAT} ) (%)</th>
</tr>
</thead>
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<tr>
<td>Lansjärvä</td>
<td>( e^0 = 1 )</td>
<td>17.15</td>
<td>3.85</td>
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<td>( e^1 = 2.72 )</td>
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<td>( e^2 = 7.39 )</td>
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<td>12.52</td>
<td>52.37</td>
<td>15.94</td>
<td>31.70</td>
</tr>
<tr>
<td></td>
<td>( e^3 = 20.09 )</td>
<td>16.85</td>
<td>4.14</td>
<td>12.71</td>
<td>34.75</td>
<td>36.13</td>
<td>29.12</td>
</tr>
<tr>
<td></td>
<td>( e^4 = 54.60 )</td>
<td>16.91</td>
<td>3.73</td>
<td>13.18</td>
<td>13.38</td>
<td>62.91</td>
<td>23.70</td>
</tr>
</tbody>
</table>

\( R \) = total runoff; \( R_s \) = surface runoff; \( R_{sb} \) = subsurface runoff; \( R_{sb,BOT} \) = bottom drainage; \( R_{sb,TOP} \) = topographic control; \( R_{sb,SAT} \) = saturation excess.
Eq. (8)), $R_{sb, TOP}$ from using the 1000 m DEM is smaller by a factor of $e^2$ compared to that from using the 100 m DEM. If the topographic index from the 1000 m DEM is employed without adjustment by Eq. (8), $R_{sb, TOP}/R_{sb}$ is decreased from 14.92% to 1.43% in Òvre Lansjärv and from 24.67% to 2.62% in Abiskojokk (Table 6).

### 6. An alternative scheme of subsurface runoff (VISA-TOP2)

We seek an alternative parameterization of the production of subsurface runoff due to saturation excess, $R_{sb,SAT}$. In this framework, the soil column saturation excess is first used to recharge the soil

![Figure 9](image-url)
layers above the water table, while the formulations of \( R_{sb, TOP} \) and \( R_{sb, BOT} \) remain the same as in VISA-TOP1. In this case, \( R_{sb, SAT} \) can still be greater than zero if the entire soil column is oversaturated. The resulting version is called VISA-TOP2.

Fig. 9 (left panel) shows the simulations from VISA-TOP2 in direct comparison with those in Fig. 4 (right panel) from VISA-TOP1. The chief difference in the two experiments is how the saturation excess is handled. With the saturation excess water recharged back to the soil layers above the water table in VISA-TOP2, the water table depth becomes shallower than that in VISA-TOP1 as expected. The saturated fraction in VISA-TOP2 almost reaches its maximum value during the intensive recharging period. Although surface runoff increases, it is too small to compensate the decrease in subsurface runoff. Therefore, VISA-TOP2 fails to capture the observed recession curve. If the decay factor increases to \( f=6 \), subsurface runoff is largely increased (Fig. 9, right panel). The soil moisture decreases and the water table falls. Judging from the simulations of streamflow, VISA-TOP2 has the best fit with \( f=6 \), while VISA-TOP1 uses \( f=4 \). This indicates that different model formulations possess different values of optimum parameters.

Like VISA-TOP1, VISA-TOP2 is strongly sensitive to \( f \). The detailed patterns are, however, dramatically different. In VISA-TOP2, the water table is confined to near the surface and the saturated fraction spreads between 0 and 35%. As \( f \) increases, the water table falls due to increased subsurface runoff. The saturated fraction also decreases as the water table depth increases. In the case of \( f=2 \), almost the entire

![Fig. 10. Time series of the modeled unrouted subsurface runoff compared with the observed streamflow, water table depth, saturated fraction for different values of \( f \) in Övre Lansjärv using VISA-TOP2.](image)
soil column is oversaturated during the late spring when snowmelt reaches its peak (around day 1230 in Fig. 10), and the oversaturated water runs off directly.

We have also tested the model’s sensitivity to the anisotropic factor \( \alpha \) in the case of \( f=6 \), for various values of \( \alpha = e^x \), where \( x = 1, 2, 3, 4 \). The simulations of unrouted subsurface runoff, water table depth, and saturated fraction are strongly sensitive to \( \alpha \). Increasing its value directly increases subsurface runoff, lowers the water table, and reduces the saturated fraction (Fig. 11). If \( x \) is less than 2, VISA-TOP2 fails to capture the observed recession curve. The case \( x=0 \) (not shown) corresponds to the original TOPMODEL formulations, but it is unable to obtain the observed recession curve, indicating this factor is necessary in VISA-TOP2. The anisotropic factor may be interpreted here to compensate the error due to use of coarse resolution DEMs. Although the topographic index from the 1000 m DEM has been adjusted using the 100 m DEM (cf. Eq. (8)), this adjustment alone may not be adequate. Indeed, Eq. (8) was derived based on the 100 and 1000 m DEMs for 50 locations in the conterminous USA (Wolock and McCabe, 2000). It is likely that this equation may need further adjustment in the Torne–Kalix basin and that this further adjustment may be accounted for by introducing the anisotropic factor.

7. Evaluation of an intermediate version (VISA-TOP3)—without topographic index

In light of the above difficulties in producing an accurate estimate of the topographic index for the

Fig. 11. Time series of the modeled unrouted subsurface runoff, water table depth, saturated fraction for different values of the anisotropic factor (\( x \) represents the power of \( \alpha = e^x \)) in Övre Lansjärv using VISA-TOP2.
global continents, it is attractive to develop a topography-related runoff parameterization which does not require the topographic index data set. In the simplest case, the topographic characteristics may be parameterized as constants for all land points, and the saturated fraction and subsurface runoffs are only determined by the soil moisture represented by the water table depth. The resulting version is called VISA-TOP3.

Surface runoff still takes the form of Eq. (2), but the saturated fraction is parameterized as

$$F_{\text{sat}} = F_{\text{max}} e^{-f z_j}$$  \hspace{1cm} (10)

where $F_{\text{max}}$ and $z_j$ are the maximum saturated fraction and the mean water table depth of a grid cell. $F_{\text{max}}$, corresponding to the cumulative distribution function of the topographic index when the water table depth is zero, is estimated here as 0.35 from Fig. 2.

Subsurface runoff due to the topographic control is parameterized as

$$R_{\text{sb, TOP}} = R_{\text{sb, max}} e^{-f z_j}$$  \hspace{1cm} (11)

where $R_{\text{sb, max}}$ is the maximum subsurface runoff and $R_{\text{sb, max}} = 1.0 \times 10^{-4}$ (mm/s) = 8.64 (mm/day). The other components of subsurface are the same as in VISA-TOP2.

Using $f = 2$ for Övre Abiskojokk and $f = 6$ for Övre Lansjärv, VISA-TOP3 also reproduces the observed streamflow (Fig. 12). Although the simulations of total runoff, water table depth, soil moisture, and subsurface runoff are similar to those from VISA-TOP2, the saturated fraction is much smaller than that from VISA-TOP2 (cf. Figs. 9 and 12), indicating that...
the simulated surface runoff is different between these two schemes. Like VISA-TOP2, VISA-TOP3 is sensitive to \( f \) (cf. Figs. 10 and 13). However, the patterns in the simulations of water table depth and saturated fraction are dramatically different, suggesting that \( f \) plays different roles in different model formulations.

8. Summary and conclusion

Three topography-based runoff schemes in VISA have been evaluated using the data set for Övre Lansjärv and Övre Abiskojokk. VISA-TOP1 differs from VISA-TOP2 only in how to treat saturation excess or oversaturated soil water from the soil layers. In VISA-TOP1, the oversaturated soil water is thrown out of the soil column; hence, it no longer plays a role in the ensuing soil water budgets. In VISA-TOP2, this oversaturated soil water is recharged back to the unsaturated soil layers above the water table; hence, it continues to involve in the water budgets. Unlike VISA-TOP1 and VISA-TOP2, VISA-TOP3 relaxes its dependence on the topographic parameters. Using a global mean constant of the topographic index, \( F_{\text{sat}} \) and \( R_{\text{sh,TOP}} \) are only dependent on the water table depth and the saturated hydraulic conductivity decay factor. The oversaturated soil water is treated the same in both VISA-TOP2 and VISA-TOP3.

The intercomparison of the modeled unrouted runoff with the observed streamflow reveals that VISA-TOP1 reproduces the daily and seasonal variations of streamflow. Sensitivity tests indicate that the simulations of runoff, water table depth, and saturated fraction are strongly sensitive to the saturated hydraulic conductivity decay factor, the anisotropic factor, lake fractions, and the topographic index.

Fig. 13. Time series of the modeled unrouted subsurface runoff compared with the observed streamflow, water table depth, saturated fraction for different values of \( f \) using VISA-TOP3.
modification due to DEM resolutions. The decay factor, $f$, controls the timing and partitioning of subsurface runoff. The anisotropic factor is necessary for the successful implementation of TOPMODEL into SWATs. The topographic index computed from coarser resolution DEMs need to be adjusted using available finer resolution DEMs. The regression equation proposed by Wolock and McCabe (2000) improved the simulations of the topography-controlled subsurface runoff.

In the VISA-TOP2 scheme, the topography-controlled subsurface runoff is dominant because the oversaturated water is recharged to upper unsaturated soil layers to raise the water table. VISA-TOP2 reproduces the simulations from VISA-TOP1 provided that different values of $f$ are used in both schemes. The water table depth and the saturated fraction in VISA-TOP1 and VISA-TOP2 show dramatically different responses to changes in $f$. Relaxing the dependence on the topographic parameters, VISA-TOP3 also is capable of reproducing the observed streamflow provided that $f$ is adjusted.

In this study, $f$ and $a$ are identified as two adjustable parameters. Both have physical meanings at local scales. In a catchment or a GCM grid cell, however, both parameters are quasi-physical and need to be calibrated against observations due to large subgrid heterogeneity of surface conditions. The anisotropic factor is a parameter which was first introduced by Chen and Kumar (2001) merely to produce the desired streamflow. It may be interpreted as a compensating agent to reduce the error caused from using a coarse-resolution DEM in calculating the topographic index. The calibrated values of $f$ are different in VISA-TOP1, VISA-TOP2 and VISA-TOP3, indicating that the calibrated parameters and the model formulations should not be separated. Such issue has been discussed by Beven (1997) in terms of the model equifinality. Additional research is required to evaluate these formulations and parameters on continental and global scales.

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References

Ducharne, A., Koster, R.D., Suarez, M.J., Stieglitz, M., Kumar, P.,


