DEM resolution effects on shallow landslide hazard and soil redistribution modelling

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Abstract

In this paper we analyse the effects of digital elevation model (DEM) resolution on the results of a model that simulates spatially explicit relative shallow landslide hazard and soil redistribution patterns and quantities. We analyse distributions of slope, specific catchment area and relative hazard for shallow landsliding for four different DEM resolutions (grid sizes of 10, 25, 50 and 100 m) of a 12 km² study area in northern New Zealand. The effect of DEM resolution is especially pronounced for the boundary conditions determining a valid hazard calculation. For coarse resolutions, the smoothing effect results in a larger area becoming classified as unconditionally stable or unstable. We apply simple empirical soil redistribution algorithms for scenarios in which all sites with a certain landslide hazard fail and generate debris flow. The lower initial number of failing cells but also the inclusion of slope (limit) in those algorithms becomes apparent with coarser resolutions. For finer resolutions, much larger amounts of soil redistribution are found, which is attributed to the more detailed landscape representation. Looking at spatial patterns of landslide erosion and sedimentation, the size of the area affected by these processes also increases with finer resolutions. In general, landslide erosion occupies larger parts of the area than deposition, although the total amounts of soil material eroded and deposited are the same. Analysis of feedback mechanisms between soil failures over time shows that finer resolutions show higher percentages of the area with an increased or decreased landslide hazard. When the extent of sites with lower and higher hazards are compared, finer grid sizes and higher landslide hazard threshold scenarios tend to increase the total extent of areas becoming more stable relative to the less stable ones. Extreme care should be taken when quantifying landslide basin sediment yield by applying simple soil redistribution formulas to DEMs with different resolutions. Rather, quantities should be interpreted as relative amounts. For studying shallow landsliding over a longer timeframe, the ‘perfect’ DEM resolution may not exist, because no resolution can possibly represent the dimensions of all different slope failures scattered in space and time. We assert that the choice of DEM resolution, possibly restricted by data availability in the first place, should always be adapted to the context of a particular type of analysis. Copyright © 2005 John Wiley & Sons, Ltd.

Keywords: digital elevation model; resolution; scale; landslide modelling; sediment yield

Introduction

Topographically-based modelling of catchment processes has become very popular in recent applied environmental research, mainly due to the advances in availability and quality of digital elevation models (DEMs) (Moore et al., 1991; Goodchild et al., 1993; Wise, 2000). Digital elevation data available from DEMs are sometimes of direct interest (e.g. in erosion, sedimentation and landscape evolution studies), but elevation values are most often used in algorithms to calculate surface derivatives such as slope, aspect, flow direction and upstream contributing area. Catchment boundaries and stream drainage networks can be derived from those topographic attributes as well. The results of these DEM analyses are used in many terrain modelling applications: distributed hydrological modelling (Beven and Moore, 1993), prediction of surface saturation zones (O’Loughlin, 1986; Barling et al., 1994), erosion-deposition models (Desmet and Govers, 1996; Schoorl et al., 2000), hillslope stability and landslide (hazard) models...
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Earth Surf. Process. Landforms
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index can be expressed as $\ln(\alpha/\tan\beta)$, where $\alpha$ is the specific catchment area and $\beta$ is the local slope. The index essentially is a measure of the tendency of water to accumulate at one position on a slope. In general, with coarsening resolution, $\beta$ tends to drop because local variation in terrain is smoothed, whereas the distribution of $\alpha$ tends to shift towards larger values (Band and Moore, 1995; Wilson et al., 2000). Several researchers compared distributions or
Influence of DEM Resolution on Digital Terrain Analysis
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mean values of $\ln(a/\tan\beta)$ computed with DEMs with differing resolutions (Quinn et al., 1991, 1995; Band et al., 1993; Chairat and Delleur, 1993; Zhang and Montgomery, 1994; Saulnier et al., 1997). In general, using coarser DEMs, a larger percentage of high index values was obtained. Wolock and Price (1994) argue that, for TOPMODEL, coarse resolution DEMs are not necessarily inappropriate because an implicit assumption is that the water table configuration mimics surface topography and may be smoother and better represented by a coarse resolution DEM. Braun et al. (1997) and Becker and Braun (1999) show that an acceptable approximation of the ‘real’ distribution of the topographical index in a catchment can be derived from a low resolution DEM by the use of simple scaling.

DEM resolution and influence on topographical attributes and modelling results

Bates et al. (1998) illustrate the interaction between DEM resolution and model grid size of a hydraulic and hydrological model. They describe DEM resolution as a first filter on topographic information content assimilated into a model. They also point to a second topographic information filter when the grid size of other parameter inputs is set at a lower resolution than that of the DEM. Gains in accuracy supposedly given by producing models that rely on more physically complex descriptions of processes, may be nullified by the lack of sufficient high resolution data. A third concern, also noted by Grayson et al. (1993), is that if the information content of the DEM drops below some threshold, certain assumptions in distributed models, particularly if flow routing is involved, may not be met.

By extracting watershed geometry from a DEM and using it in an event-based distributed model for calculating surface runoff, Thieken et al. (1999) found that flow path lengths, drainage density and time to peak flow decreased whereas peak flow rate, maximum total flow length and runoff volume increased with decreasing DEM resolution.

Wilson et al. (2000), using TAPES-G for deriving primary topographic attributes, found decreasing slope gradients, more short flow paths (measured in terms of number of cells) and increasing specific catchment area with increasing cell size. They also delineated channel/net deposition and net erosion cells by the change in sediment transport capacity index (Moore and Burch, 1986) across a grid cell. For the net erosion areas, they also describe the sensitivity to DEM resolution for the Revised Universal Soil Loss Equation length-slope (RUSLE-LS) factor, using flowpath length, and the Moore/Wilson (MW) sediment transport capacity index, using specific catchment area (Moore and Wilson, 1992). The two variables increased with increasing grid size and different statistical and spatial distributions were generated. Furthermore, fundamentally different values were obtained and this is likely to cause additional problems when substituting one for the other in empirical models (see also Mitasova et al., 1996; Desmet and Govers, 1997).

Applying a simple single-process model for erosion and sedimentation on two artificial DEMs with five different resolutions, Schoorl et al. (2000) described an artificial mathematical overestimation of erosion and a realistic natural modelling effect of underestimating resedimentation with coarsening resolution. The first effect could be handled by introducing a systematic correction factor, the second by modelling in a multi-scale framework and using resedimentation rates from finer resolutions in simulations for larger areas with coarser resolution.

Thompson et al. (2001) statistically and visually compared terrain attributes and quantitative soil–landscape models derived from DEMs with different horizontal resolution and vertical precision. They suggest that the vertical precision must increase with increasing horizontal resolution so that it remains greater than the average difference in elevation between grid points in the DEM. Furthermore, they found similar capabilities in predicting A horizon depth for 10 m and 30 m resolution DEMs and therefore state that higher-resolution DEMs may not be necessary for generating useful soil–landscape models.

In general, the significance of shifts in topographical attributes derived from DEMs with different resolution needs to be assessed relative to the sensitivity of the models using this information and landscape features of interest should guide the choice of resolution (Band and Moore, 1995). Zhang and Montgomery (1994) found that runoff processes in their area were controlled by physical properties of the landscape of about 10 m and suggest a 10 m grid size as a rational compromise between increasing resolution and data volume for simulating geomorphic and hydrological processes. Garbrecht and Martz (1994) found that, for extracting drainage properties from their DEM, the grid area should be less than 5 per cent of the network reference area to reproduce important drainage features.

DEM resolution and landslide modelling

Models used to delineate the location and calculate the potential for shallow landsliding in a grid-based DEM environment are being used with a variety of grid sizes, ranging from as coarse as 4-05 hectares (Ward, 1981) to a few metres (e.g. 2 m; Dietrich and Montgomery, 1998). In general, cell size is selected by the user and should depend on the quality and density of the input data, size of the area to be mapped and accuracy required for the output data (Ward, 1981). Borga et al. (1998, 2002) use 10 m grids in one basin, ‘consistent with the resolution of the original topographic data and the low grid-to-hillslope length ratio’, and 5 m grids in another catchment ‘to portray individual
landslides accurately’. Burton and Bathurst (1998) use a dual-resolution approach, modelling basin hydrology at a 200 m resolution, for time and memory requirements, and predicting landslide hazard and potential sediment yield at 20 m resolution, which they consider to be appropriate for landslide simulation. Vanacker et al. (2003) use a 5 m grid size DEM in a factor of safety analysis, a resolution judged suitable for studying the linkage between land-use change and slope stability.

Few studies have been done on the identification and quantification of the influence of DEM resolution on landslide hazard assessment and the resulting soil redistribution pattern. Dietrich and Montgomery (1998) give two examples of how the distribution of the hydrologic ratio, expressed as log(q/T) (q is steady-state rainfall [m day⁻¹], T is transmissivity [m² day⁻¹]), varies with different DEM resolutions in a landscape. They also show a comparison for a shallow landslide location and hazard prediction (with SHALSTAB) between a 30 m and a 6 m DEM and a 10 m and 2 m DEM. For both cases they conclude that, although percentages of the landscape in the moderate landslide hazard classes are similar for coarse and fine resolutions, the spatial patterns in general differ in important ways. In the finer resolution case, patterns of relative slope stability are much more strongly defined by local ridge and valley topography. Low log(q/T) values (meaning a higher slope instability potential) are more concentrated in steep valleys, rather than spread out across the landscape. They conclude that with finer resolution topography, sites with highest instability increase and can be delineated more precisely, rather than mapping broad zones of instability in the case of coarser resolutions.

Materials and Methods

For studying long-term landscape dynamics in a study area in northern New Zealand, where shallow landsliding is the dominant erosional process, we developed a landslide model component LAPSUS-LS (Claessens et al., in press). This module is embedded in the LAPSUS modelling framework. LAPSUS (landscape process modelling at multi dimensions and scales; Schoorl et al., 2000) originally addresses on-site and off-site effects of current and possible soil redistribution by water runoff and tillage erosion. Landscape evolution within a timeframe of decades is simulated in a dynamic way by adapting digital elevation data between yearly timesteps, according to the calculated soil redistribution. (Schoorl and Veldkamp, 2001; Schoorl et al., 2002).

The overall aim of LAPSUS-LS is to assess the impacts of shallow landsliding on long-term landscape dynamics and to identify possible feedbacks with other hillslope processes. LAPSUS-LS works on a time scale of years to decades; it is not the intention to simulate detailed changes in channel geometry or hillslope geomorphology caused by individual failures. In spite of its assumptions and limitations, this approach has been demonstrated to retain the essence of the physical controls of topography and soil properties on landsliding and remains parametrically simple for ease of calibration, validation and application (Claessens et al., in press).

LAPSUS-LS is a combination of several modelling steps. First it calculates relative landslide hazard distribution from topographical and geotechnical attributes. Historical rainfall–landslide distribution datasets and magnitude–frequency scenarios are then used to calibrate and run the model. Soil redistribution algorithms are applied to quantify and visualize feedbacks between mass movements or interactions with other geomorphic processes.

Relative hazard for shallow landsliding

For the analysis of shallow landslide hazard we combine a steady-state hydrologic model with a deterministic infinite slope stability model to delineate areas prone to landsliding due to surface topographic effects on hydrologic response. This method has been described by Montgomery and Dietrich (1994) and is based on earlier formulations proposed by O’Loughlin (1986). Over the last decade, this approach and adaptations of it have been proven to be very practical and perform well in many applications (Dietrich et al., 1995; Dietrich and Montgomery, 1998; Wu and Sidle, 1995; Borgia et al., 1998, 2002; Montgomery et al., 2000; Duan and Grant, 2000; Pack et al., 2001; Vanacker et al., 2003; Fernandes et al., 2004). Our approach is based on the original method but has some differences:

1. The cohesion term is retained in the infinite slope stability model to account for soil cohesion and additional strength by root reinforcement.
2. We use grid-based rather than contour-based DEM methodology.
3. Unlike some researchers who account for variability and uncertainty of terrain attributes in more probabilistic approaches (Dietrich et al., 1995; Duan and Grant, 2000; Pack et al., 2001; Zaitchik et al., 2003), we restrict ourselves to a deterministic approach, lumping all but the topographic parameters in the watershed, in this way focusing solely on DEM resolution effects.
We do not use the factor of safety directly as in some applications, but derive the minimum steady-state rainfall predicted to cause slope failure.

We call $Q_\text{cr}$ [m day$^{-1}$] the critical rainfall, which can be written as:

$$Q_\text{cr} = T \sin \theta \left( \frac{b}{a} \right) \left( \frac{\rho_s}{\rho_w} \right) \left[ 1 - \frac{(\sin \theta - C) \cos \phi}{(\cos \theta \tan \phi)} \right]$$

(1)

where $T$ is saturated soil transmissivity [m$^2$ day$^{-1}$], $\theta$ is local slope angle [°], $a$ the upslope contributing area [m$^2$], $b$ the unit contour length (in our grid-based approach the grid resolution [m] is taken as the effective contour length as in Pack et al., 2001), $\rho_s$ wet soil bulk density [g cm$^{-3}$], $\rho_w$ the density of water [g cm$^{-3}$] and $\phi$ the effective angle of internal friction [°]. $C$ is the combined cohesion term [-], made dimensionless relative to the perpendicular soil thickness and defined as:

$$C = \frac{C_r + C_s}{h \rho g}$$

(2)

where $C_r$ is root cohesion [N m$^{-2}$], $C_s$ soil cohesion [N m$^{-2}$], $h$ perpendicular soil thickness [m] and $g$ the gravitational acceleration constant (9.81 m s$^{-2}$). With the assumptions and boundary conditions used in deriving Equation 1 (see Claessens et al., in press), we can express the conditions for upper and lower thresholds for elements that can possibly fail. Unconditionally stable areas are predicted to be stable, even when saturated and satisfy

$$\tan \theta \leq \left( \frac{C}{\cos \theta} \right) + \left( 1 - \frac{\rho_w}{\rho_s} \right) \tan \phi$$

(3)

Unconditionally unstable elements, in most cases bedrock outcrops, are unstable even when dry and satisfy

$$\tan \theta > \tan \phi + \left( \frac{C}{\cos \theta} \right)$$

(4)

In the model, slope and contributing area are calculated at each grid point and the other parameters are lumped for grid cells within an area of same parent material and vegetation type. In this way the spatial distribution of critical rainfall values can be calculated expressing the relative potential for shallow landslide initiation.

A sensitivity plot with the relative importance of each variable in the calculation of $Q_\text{cr}$ [m day$^{-1}$] following Equation 1 is given in Figure 1. This graph shows the relative change of the critical rainfall value as a function of the relative divergence of each input variable from a given set of default input values. Obviously, $T$, $\rho_s$ and $C$ show a linear, positive correlation with $Q_\text{cr}$. Increasing specific catchment area or slope lowers $Q_\text{cr}$; higher internal friction angles yield a lower landslide hazard. Given the set of default input parameters, the negative correlation of slope with $Q_\text{cr}$ is larger than that of specific catchment area. Internal friction has the largest positive correlation with $Q_\text{cr}$. Hence, in general it is the most effective in lowering relative landslide hazard with increasing input values.

**Failed slope material redistribution**

For determination of debris flow trajectories, sophisticated formulations applicable to well-specified individual debris flows were inappropriate, and simpler empirical formulas were developed (see Claessens et al., in press). In the erosional phase, the debris flow is supposed to erode an amount of unstable soil material $S$ [m] on its way following the steepest descent direction and estimated as:

$$S = \frac{\rho_w \cos \theta (\tan \theta - \tan \alpha) a}{C_s}$$

(5)

with $\alpha$ [°] the minimum local slope for debris flow movement and $a$ [m$^2$] a dimensional correction factor.

The point at which deposition begins or where the addition of eroded material within the erosional phase reduces to zero is reached once the gradient falls below a certain slope angle $\alpha$. We use the elevation loss within the erosional phase as a measure of debris flow momentum at the start of deposition. This is adapted from a method described by Burton and Bathurst (1998) to quantify percentage landslide sediment delivery to streams. The number
Figure 1. Sensitivity plot for critical rainfall (Equation 1), showing the percentage change of its value as a function of the divergence of each input variable from a given set of default input values.

of downslope grid cells involved in the deposition of debris flow material, defined as ‘cell-distance’ $D \ [\text{--}]$, is calculated as:

$$D = \left( \frac{\Delta y \phi}{b} \right)$$

where $b$ is the grid resolution [m], $\Delta y$ [m] is the elevation difference between the head of the slide and the point at which deposition begins, and $\phi \ [\text{--}]$ is an empirically derived fraction set at 0·4 (Vandre, 1985; Burton and Bathurst, 1998). To incorporate importance of hillslope morphology for the spatial deposition pattern, the accumulated soil material is routed with multiple flow principles: for downslope neighbours of the point where deposition starts we express the sediment, which is effectively delivered to grid cell $i$, as:

$$S_i = \frac{B_{i-1}}{D_{i-1}} f_i$$

The term $B_{i-1}/D_{i-1}$ is the amount of sediment coming from the erosion part upslope (grid cell $i - 1$), divided by the cell-distance (Equation 6), and deposited in grid cell $i$. The remaining sediment budget of grid cell $i$, which is not deposited but ‘passed through’ to grid cell $i + 1$, can be expressed as

$$B_i = B_{i-1} \left( 1 - \frac{1}{D_{i-1}} \right) f_i$$

In Equations 7 and 8, $f_i$ is the fraction of sediment ($S$) or sediment budget ($B$) allocated to each lower neighbour and determined by the multiple flow concept by Quinn et al. (1991). Sediment and sediment budget are divided to all downslope neighbours, using a slope-dependent weighting factor for each fraction:

$$f_i = \frac{(\Lambda)^{f_i}}{\sum_{j=1}^{8} (\Lambda)^{f_j}}$$

where fraction $f_i$ of the total amount out of a cell in direction $i$, is equal to the slope gradient (tangent) $\Lambda$ in direction $i$ powered by factor $p$, divided by the summation of $\Lambda$ for all (never more than eight) downslope neighbours $j$ powered by factor $p$. In each downslope grid step, the cell-distance is lowered by one and when $D < 1$ all the remaining sediment is deposited and the debris flow halts.

With the aim of developing a model that does not require the rheological properties of debris flow, this empirical methodology can easily incorporate a slope limit on debris flow movement. It takes into account hillslope morphology for erosion and deposition and is based on elevation differences, thus convenient when topography is represented by a digital elevation model. More details and discussion about the limitations of the model can be found in Claessens et al. (in press).

**Study area**

The Waitakere Ranges Regional Park lies immediately west of Auckland, New Zealand (174.8° E, 36.9° S, Figure 2). Altitude ranges from sea level to 474 m. The area has a warm and humid climate with a mean annual rainfall ranging from about 1400 mm near the Tasman coast up to 2030 mm at the higher altitudes (Auckland Regional Council, 2002). Much of the area is still covered with dense rainforest or regenerating native vegetation. The rugged topography of the area is caused by the resistant nature of the component rocks which are predominantly Miocene volcanic rocks. The landscape is mantled by very deep soils but locally bedrock crops out on steep slopes, in cliffs, and volcanic intrusions. Soils are classified as haplic Acrisols according to the FAO classification (FAO, 2001). The clay composition of the soils is dominated by kaolinite but varying fractions of smectite, halloysite and vermiculite clays are present. The study area chosen for this analysis is the c. 3 km by 4 km catchment of the Piha and Glen Esk streams and their tributaries (Topographic Map of New Zealand 260-Q11 and Pt. R11, 416 728–458 697). The geology of this part mainly consists of Miocene volcanic breccia and conglomerate (Piha Formation; Hayward, 1976). Slope failure is very common involving 1 to 40 m$^3$ of soil material on slopes steeper than 18° in cleared areas and on slightly steeper slopes in bush-covered areas (Hayward, 1983).

![Figure 2. Location of the study area in the North Island of New Zealand.](image-url)
Parameterization

Although we do have reliable field and laboratory measurements of $T$, $C$, $\rho$, and $\phi$ for the different parent materials of the study area and could treat these parameters as spatially distributed (Claessens et al., in press), for this application all variables, except slope and contributing area, are lumped in the watershed. Lumping of parameters has also been done by Barling et al. (1994), Dietrich and Montgomery (1998) and Wu and Sidle (1995), due to lack of sufficient measured catchment properties. Lumped input parameters are estimated on the basis of field and laboratory measurements (Claessens et al., in press), and are 15 m$^2$ day$^{-1}$ for $T$, 0.2 for $C$, 1.5 g cm$^{-3}$ for $\rho$, and 35° for $\phi$. Exponent $p$ of Equation 9 for the determination of the fraction for multiple flow directions is set at 4 (Holmgren, 1994).

Terrain analysis: DEM, slope and specific catchment area

To investigate the effect of DEM resolution, 100 m, 50 m and 25 m DEMs were aggregated in ArcView GIS from the same 10 m DEM. The 10 m DEM in turn was derived by TIN interpolation from 10 m elevation contours electronically created from 3D stereo photographs. Although aware of the possible effects of the choice of slope and flow routing algorithms briefly discussed above, we will not explore these effects further here and focus solely on the effect of DEM resolution.

For calculation of slope, the local cell-to-cell slope is used, rather than using a smoothing multiple cell window, as done in most GIS procedures. The upslope contributing drainage area is calculated using the concept of multiple flow by Quinn et al. (1991), to represent the convergence or divergence of flow under topographic control. Although Tarboton (1997) correctly points out that this method is ‘dispersive’, Dietrich and Montgomery (1998) discuss that it avoids grid artefacts and gives reasonable estimates for the total drainage area for each cell, when used in Equation 1. Furthermore, we also take grid cell size as effective contour length $b$ in the definition of specific catchment area because other algorithms that calculate $b$ based on the sum of flow directions will be influenced by the orientation of the topography relative to the grid system (Dietrich and Montgomery, 1998).

Results and Discussion

Influence of grid size on slope, specific catchment area and critical rainfall

To explore how different DEM resolutions influence relative shallow landslide hazard distribution, the behaviour of the two topographical attributes derived from the DEM and used in the hazard assessment, i.e. local slope and contributing area, is analysed. These parameters constitute a topographical and hydrological characterization of the study area, respectively.

Figure 3 shows the cumulative distribution of slope values within the study area for the four DEM resolutions. Although the trends are quite similar (note the same near absence of slopes around 0.20 rad for the four resolutions), it is clear that the coarser resolutions show a larger contribution of lower slope angles and fewer short steep slopes. Both effects become less pronounced towards the finest grid size, reflecting the progressively smaller improvement made by representing the original 10 m contour data of our area with increasingly smaller grid cells. It must be stressed that the method of slope computation is very important when analysing these resolution effects. As mentioned, we use a cell-to-cell slope computation, meaning that slope is calculated over a ‘slope distance’ equal to the DEM resolution, i.e. ranging from 10 to 100 m with coarsening resolution. When using multiple cell windows to compute slope, the ‘slope distance’ will increase and the smoothing effect of coarser resolutions will be even more pronounced. The ‘yardstick’ (Mandelbrot, 1986) used to compute a slope should not automatically depend on the DEM resolution but rather on the process to be simulated.

The cumulative distribution for the computed specific catchment area (contributing area per unit contour length) for the different resolutions is plotted in Figure 4. The minimum values are directly related to the grid size and are equal to the area of one cell divided by the flow width or contour length. In other words, they equal the DEM resolution. Coarser DEM resolutions also have a higher contribution of high specific catchment area values to the distribution: progressively more drainage area tends to accumulate when directing flow downslope towards the outlet cell due to the higher unit contributing area for the larger grid sizes. In the end, the value of contributing area for the outlet cell of the watershed is approximately the same for the four resolutions. Similar smoothing effects of coarser resolutions on derived topographic attributes were found by several researchers for various landscapes (Quinn et al., 1991, 1995; Wolock and Price, 1994; Zhang and Montgomery, 1994; Moore, 1996; Wilson et al., 2000).
The effects of resolution on the cumulative distributions of slope and specific catchment area have a direct impact on the calculated values for the critical steady-state rainfall for landslide initiation $Q_{cr}$ [m day$^{-1}$] (Equation 1). Figure 5 shows, for the different DEM resolutions, the percentage of the study area having a certain critical rainfall value, interpreted as the relative potential for landsliding or becoming ‘unstable’. The most pronounced effect is the influence of slope distribution on the boundary conditions for using Equation 1, i.e. assigning a critical rainfall value to a grid cell. The intercept with the Y-axis or area with ‘zero’ critical rainfall value equals the amount of unconditionally unstable cells according to Equation 2. In our application, this condition is only influenced by the local slope as a variable factor and more unconditionally unstable cells are encountered with more terrain variance and steeper slopes. The other boundary condition (Equation 3), excluding unconditionally stable areas from landslide hazard calculation, is also completely dependent on local slope distribution and for the same reason, less unconditionally stable cells are found with finer resolutions. In general, the effect of smoothing or filtering of topographic information is very clear and the detail obtained in the lower critical rainfall value ranges is progressively lost with coarsening resolution. Within the range of valid critical rainfall calculations according to Equation 1, besides local slope, differences in specific catchment area distribution also become important. This is more difficult to interpret from Figure 5 but the effect of the coarse resolution DEM having more high specific catchment area values, possibly lowering $Q_{cr}$, is mostly
nullified because those higher values tend to occur in the valley bottom towards the outlet cell, where slopes are generally low enough for classification as unconditionally stable. This is confirmed when comparing specific catchment area descriptive statistics, split up between areas having a possible valid critical rainfall calculation and unconditionally stable parts (Table I).

### Table I. Descriptive statistics for specific catchment area and slope values for the different DEM resolutions

<table>
<thead>
<tr>
<th>Grid size (m)</th>
<th>Landslide hazard</th>
<th>Specific catchment area (m²)</th>
<th>Slope (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Min.</td>
<td>Max.</td>
</tr>
<tr>
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<td></td>
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<td>100·0</td>
<td>23·074·8</td>
</tr>
</tbody>
</table>

### Influence of grid size on landslide soil redistribution quantities and spatial patterns

After calculating the relative hazard for shallow landsliding \( Q_r \) over the catchment, the interactions between possible initial collapse and resulting debris flow soil redistribution can be assessed by applying Equations 5 to 8 as described above. DEM elevation values can then be adapted according to the amounts [m] of landslide erosion (Equation 5) and deposition (Equation 7). When, for a next model run or timestep, \( Q_r \) is calculated again with the adapted DEM, it is possible to visualize feedbacks between mass movements or interactions with other geomorphic processes by comparing landslide hazard maps between timesteps. Historical rainfall–landslide distribution datasets can provide the link between real rainfall events (with their magnitude and frequency), triggering slope failure at certain locations over time and the critical steady-state rainfall value assigned to those cells. In this way, the time frame represented by one...
timestep or model run can be quantified. Alternatively, since those long-term and spatially explicit datasets are rarely available, one timestep can be interpreted as a scenario with the average frequency of many rainfall events of different intensity and duration that trigger shallow landslides over years translated in the failure of all sites with a critical steady-state rainfall smaller than a certain value. Figure 6 shows, for a small part of the study area, the landslide soil redistribution pattern resulting from a scenario in which all cells with critical rainfall of 0·2 m day\(^{-1}\) or less are triggered. Zones with increased or decreased landslide hazard for the next timestep are also delineated after comparing the initial and resulting critical rainfall map.

After running the modelling steps for a certain scenario, the total amounts [m\(^3\)] of eroded and deposited soil material can be quantified. These quantities are the same because consecutive fluvial sediment transport out of the catchment is not yet accounted for in the model. Figure 7 gives a comparison of total amounts of erosion for four different scenarios applied to the four DEMs. The maximal \(Q_{cr}\) value in the catchment is 0·77 m day\(^{-1}\). The corresponding scenario is a measure of total possible landslide erosion/sedimentation within the catchment or the most extreme case in which all sites with a valid hazard value are triggered. The other scenarios with progressively lower maximal \(Q_{cr}\) values represent more realistic situations in which only areas with consecutive higher hazard values are affected. The influence of grid size on critical rainfall distribution (Figure 5) has a direct impact on the amount of sites triggered and the resulting erosion/sedimentation quantities in each scenario. The maximum possible quantities are already reached for lower threshold critical rainfall scenarios when resolution is coarsened and again the smoothing of the landscape becomes apparent. Furthermore because of this, the effect of a higher unit area for the coarser resolutions, used to calculate the amount of soil material in m\(^3\), is not enough to counteract the diversion of the amounts for smaller grid sizes towards much higher values with increasing critical rainfall threshold.

Bearing in mind this resolution effect, extreme care should be taken when computing, for example, quantitative basin sediment yields with grid-based soil redistribution algorithms. In particular, the quantification of landslide

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**Figure 6.** Detail of landslide soil redistribution pattern and changes in landslide hazard between timesteps.

**Figure 7.** Total amounts of landslide erosion for different critical rainfall thresholds and DEM resolutions.
erosion and deposition, which involves rather scattered processes within a catchment compared to the more spatially continuous process of water (surface) erosion, is extremely sensitive to the resolution chosen to represent the dimensions of landslides and debris flows typical of a study area. One could even argue that it is impossible to choose one single best resolution to represent the characteristics of all failures occurring in a landscape over time and to quantify their soil redistribution accurately with simple empirical algorithms.

Since the model was not developed for calculating erosion or sediment yields in a quantitative way but for studying spatial patterns of long-term soil redistribution and feedback mechanisms, the impact of resolution on these patterns is assessed. In Figure 8, the percentages of the total area affected by erosion and deposition are plotted for different scenarios and DEM resolutions. Almost the same patterns as for the calculated quantities (Figure 7) are visible. Again, the larger unit area for coarser resolutions and thus the larger possible contribution to the percentage of the area affected is not enough to counteract the effect of many more possibly triggered landslide sites and resulting soil redistribution with finer resolutions. Except for the 100 m DEM, larger parts of the area tend to be involved in landslide erosion than deposition (Figure 9), although the amounts of soil material eroded and deposited are the same.

![Figure 8](image1.png)

**Figure 8.** Percentages of the total area affected by erosion and deposition for different critical rainfall thresholds and DEM resolutions.

![Figure 9](image2.png)

**Figure 9.** Comparison of percentages of the total area affected by landslide deposition and erosion for different critical rainfall thresholds and DEM resolutions.
Here, the higher unit areas for coarser resolutions do relatively raise the extent of deposition area because of the multiple flow principles used in the sedimentation algorithm. These trends are of course also dependent on the choice of $\phi$ [−], the empirically derived fraction of the elevation difference between the head of the slide and the point at which deposition begins, set at 0.4 in the deposition algorithm (Equation 6).

Figure 10 shows percentages of the area becoming relatively less or more stable after comparison of landslide hazard maps before and after a model run scenario with different threshold $Q_{cr}$ values. For finer resolutions, not only are more cells initiated according to the scenario, i.e. having a critical rainfall of the scenario threshold or less, but also the resulting soil redistribution pattern is more detailed. After comparing initial and final hazard maps, more variation is encountered with finer resolutions and higher percentages of the area have a changed (higher and lower) landslide hazard. When the extents of areas with lower and higher hazard are compared (Figure 11), finer resolutions and lower critical rainfall thresholds tend to increase the total extent of areas becoming more stable relative to the less stable ones. Finer resolution DEMs explain more terrain variance and seem in this way more effective in lowering the

**Figure 10.** Percentages of the area getting a lower or higher landslide hazard after a model run scenario with different critical rainfall thresholds and DEM resolutions.

**Figure 11.** Comparison of percentages of the total area getting a lower and higher landslide hazard for different critical rainfall thresholds and DEM resolutions.
overall modelled landslide potential; this effect is more strongly pronounced when dealing with lower critical rainfall thresholds.

Conclusions

Coarsening resolution has a smoothing effect on landscape topographical representation. Furthermore, the effect of DEM resolution on these attributes has a major impact on the distribution of the critical steady-state rainfall value, interpreted as a relative hazard for shallow landsliding. For coarser resolutions, the smoothing effect results in a larger area being excluded from the valid critical rainfall range and those parts are classified as unconditionally stable or unstable. When applying simple empirical soil redistribution algorithms, not only the lower initial amount of failing cells with coarser resolution but also the inclusion of slope (limit) in the algorithms becomes apparent. Although the model was not developed specifically to address quantitative landslide erosion and sedimentation rates, the amounts of eroded and deposited material were calculated for different scenarios and DEM resolutions. For smaller cell sizes, much higher amounts of soil redistribution were found, all caused by the much more detailed landscape representation. Looking at spatial patterns of landslide erosion/sedimentation, the percentages of the area being affected by erosion or sedimentation show a similar trend. Also larger parts of the area tend to be involved in landslide erosion than deposition in general, although the total quantities of soil material eroded and deposited are the same. When comparing landslide hazard maps between model scenario runs to visualize feedback mechanisms between soil failures over time, more variation is encountered with finer resolutions, and higher percentages of the area have an increased or decreased landslide hazard. When the percentages of the area of sites with lower and higher hazard are compared, finer resolutions and lower critical rainfall thresholds tend to increase the total extent of areas becoming more stable.

As a general conclusion we can state that extreme care should be taken when quantifying landslide erosion or sediment yield by applying simple soil redistribution formulas to DEMs with different resolutions. Calculation of erosion and sedimentation quantities can still be useful but these should rather be interpreted as relative amounts. For evaluation of shallow landslide distribution and the effect on landscape dynamics over long terms, the ‘perfect’ DEM resolution will not exist because no resolution can, even for a small study area, represent the dimensions of all possible slope failures scattered in space and time. The choice of DEM resolution may be restricted by data availability in the first place but should always be done in the context of a particular type of analysis. For example, when long-term landslide hazards and soil redistribution patterns are used to explain vegetation patterns, working at the scale of a single tree, even the 10 m DEM will probably be too coarse, whereas for broader vegetation classes, 50 m cells will possibly give a reasonable result. A coarse resolution DEM may appear to be of very poor quality, but if it can produce good results, then its quality is certainly satisfactory for that particular application. Furthermore it must be stressed that the topographical and hydrological characteristics of different landscapes can vary quite fundamentally. The choice of DEM resolution should always be made against the background of the distribution of these attributes within a study area and a sensitivity analysis of the model using them. Ideally, a DEM should represent the topographical and hydrological properties derived from it in such a way that neglecting features which are possibly ‘filtered out’ does not harm the quality of the model outcome. DEM resolution is context dependent and in the example of this paper, creates a threshold of (in)sensitivity to shallow landslide hazard calculation. Assessing these sensitivities forms an essential part of good modelling practice.

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References

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