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**A network index based version of TOPMODEL for use with high resolution digital topographic data**

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**Abstract**

This paper describes the preliminary development of a network index approach to modify and to extend the classic TOPMODEL. Application of the basic Beven and Kirkby (1979) form of TOPMODEL to high resolution (2.0 m) laser altimetric data (based upon the U.K. Environment Agency's Light Detection and Ranging (LIDAR) system) to a 13.8 km$^2$ catchment in an upland environment identified large areas of saturated areas that remained unconnected from the drainage network even during an extreme flood event. This is shown to be a particular problem with using high resolution topographic data especially over large spatial scales. To deal with the hydrological consequences of disconnected areas, we present a simple network index modification in which saturated areas only connect when the topographic index is sufficient for there to be zero or negative saturation deficits along a complete flow path. This is combined with an enhanced method for dealing with the problem of pits and hollows which is shown to become more acute with higher resolution topographic data. The paper concludes by noting the implications of the research as presented for both methodological and substantive research that is currently under way.

**Keywords**

TOPMODEL, LIDAR, hydrological similarity, digital elevation models

## Introduction

A major theme in the application of semi-distributed hydrological models such as TOPMODEL is the strong sensitivity of model predictions to the quality and resolution of topographic data that are used to derive them (e.g. Bruneau *et al.*, 1995; Quinn *et al.*, 1991, 1997; Wolock and Price, 1994; Zhang and Montgomery, 1994). There has been a particular focus upon DEM resolution. This affects the distribution of both upslope contributing areas ($a$) (e.g. Bruneau *et al.*, 1995; Zhang and Montgomery, 1994) and surface slopes ($\tan\beta$) (Bruneau *et al.*, 1995; Zhang and Montgomery, 1994). Hence, the distribution of the topographic index, $\ln(a/\tan\beta)$ (Bruneau *et al.*, 1995; Quinn *et al.*, 1997; Wolock and Price, 1994; Zhang and Montgomery, 1994), patterns of saturation (Zhang and Montgomery, 1994) and drainage patterns

(Wolock and Price, 1994) are also affected. It has also been shown that: (1) the optimal form of flow routing algorithm depends upon DEM resolution (e.g. Quinn *et al.*, 1997); (2) the importance of elevation corrections in relation to pits and artificial barriers to drainage is related to grid size (Bruneau *et al.*, 1995) and there is a strong association between model behaviour, parameterisation and DEM resolution (e.g. Zhang and Montgomery, 1994; Bruneau *et al.*, 1995). Concurrently, the ease with which we can acquire high quality topographic data is increasing significantly (Lane and Chandler, 2003) and it has proven possible to obtain 1.0 m resolution digital elevation models of catchments with an elevation precision of 0.17 m using 1:3000 scale digital photogrammetry (Lane *et al.*, 2000). The recent expansion of airborne altimetry (e.g. the U.K.'s light detection and ranging or LIDAR system) as a means of acquiring high density topographic data is opening up even greater possibilities as digital elevation models, including both digital ground models and vegetation maps (e.g. Cobby *et al.*, 2001), can be obtained if both the first and the last signal returns are recorded.

In this paper, we aim to demonstrate the application of TOPMODEL to high resolution altimetric data for an upland catchment in the Yorkshire Dales National Park, U.K. The model predicts the widespread existence of disconnected saturated zones that expand within an individual storm event but which do not necessarily connect with the drainage network. To deal with this, we present a network index based modification to TOPMODEL that allows saturated zones to contribute to overland flow only if every pixel on the flow path followed to the channel network is fully saturated. Central to this is the theme identified by Beven and Freer (2001) who, whilst recognising the need to represent the connectivity of downslope saturated flows, also note that a compromise must be reached between the needs for improved and more flexible descriptions of processes deemed as important and the needs of parameter calibration.

**Application of the basic form of TOPMODEL to high resolution topographic data**

The field-site that is the subject of this work is the Oughtershaw Beck sub-catchment of the River Wharfe, North Yorkshire (Figure 1). This is an area of 13.8 $km^2$ ranging in elevation from 353 m above Ordnance Datum (OD) at the sub-catchment outlet to 640 m on the divide to the north. In this application, we begin by applying the basic form of TOPMODEL (e.g. Beven and Kirkby, 1979) to 2.0 m LIDAR data flown specially by the Environment Agency in connection with this research project. These data have been mapped from irregular LIDAR returns onto a regular 2.0 m DEM (Figure 1) and then subject to textural filtering in order to identify blunders and to remove buildings and walls. In practice, these covered a negligible part of the study area (there are three farms and only four barns in the entire sub-catchment). The topographic data were validated using independently acquired real-time kinematic GPS data with an elevation precision of ±0.03 m. Data were collected from the full range of slope angles present in the field site. Using standard error propagation (e.g. Lane *et al.*, 2003), the residual precision of the LIDAR elevations was found to be ±0.12 m.

The basis of TOPMODEL is well rehearsed (e.g. Beven and Kirkby, 1979;Beven, 1997; Beven and Freer, 2001), and so the following section is brief. TOPMODEL seeks to partition rainfall between three components: (1) overland flow ($Q_o$); (2) recharge of the unsaturated zone; and (3) flow in the saturated zone ($Q_b$). In simple terms, rain that falls on a unit of the landscape is assumed to go into storage in the unsaturated zone. If the soil is saturated, there is no recharge and the rainfall enters the channel network as overland flow, with an appropriate delay function (Beven and Kirkby, 1979). There is also flow within the saturated zone, which is estimated making two important assumptions (Beven, 1998, 2000):

(1) flow in the saturated zone can be represented by a series of steady-state approximations of the saturated zone of area $a$ (the upslope contributing area) that drains to a length of contour on the hillslope; and

(2) the hydraulic gradient within the saturated zone is approximated by the local topographic slope, $\tan\beta$, requiring topographic data of sufficient resolution to allow an adequate description of the flow pathways without violating the assumption of local parallelism of the water table and soil surface (Saulnier, *et al.*, 1997).

There are a number of different expressions for flow in the saturated zone (e.g. Beven, 1997), based upon different assumptions of the soil transmissivity function. For instance, a common assumption is that the soil transmissivity function is an exponential function of storage deficit (Beven and Kirkby, 1979), of a shape controlled by a 'soil parameter' $m$, and a transmissivity ($T_o$) at saturation (i.e. with zero deficit when the soil is saturated to the surface). Under this scenario, the local propensity to saturation is controlled by the topographic index: $\ln(a/\tan\beta)$. It is then possible to determine the saturated zone flux or base flow contribution ($Q_b$, mhr$^{-1}$) for each sub-unit of the catchment as well as the rate of recharge to the saturated zone from the unsaturated zone ($Q_v$) (e.g. Beven and Wood, 1983). Within this system, moisture accounting is treated in a lumped fashion for hydrologically similar areas: $Q_b$ and $Q_v$ are calculated for each time step and then, after combination with $Q_o$, and in order to account for all rain that falls on a given catchment, the average catchment storage deficit ($\overline{D_t}$) is updated:

$$\overline{D_t} = \overline{D}_{t-1} + \frac{\Delta t}{A}\left(Q_b - Q_v\right)_{t-1} \qquad [1]$$

Although [1] is a lumped accounting model, for a given average catchment storage deficit it is possible to determine the critical value of the topographic index above which a location within the catchment will be saturated. Thus it is possible to map the lumped predictions of storage deficit back onto a distributed map of locations where the saturation deficit is locally zero or negative and overland flow is likely to be occurring.

We began by applying this version of TOPMODEL to the LIDAR data obtained for the Oughtershaw catchment, with the soil parameter $m$ = 0.003 m and the saturated transmissivity, $T_o$ = 1.0 $m^2s^{-1}$. Figure 2a shows predicted patterns of zero saturation deficit at two time periods within a storm event (Figure 3) in June 2000. These diagrams show large areas of the catchment where the saturation deficit is zero that are not connected to the channel network by links that have zero or negative saturation deficit throughout. As the rain storm progresses, the number and extent of these areas increases. Simultaneously, the zero/negative saturation deficit areas also expand but, even then, some locations remain unconnected. Figure 2c shows the associated topographic index map. Low values of the topographic index will lead to the disconnection shown in Figures 2a and 2b. It is clear that low values are found on steep slopes where even though the upslope contributing area may be high, the high slope results in a low value of the topographic index, and less likelihood that saturation will develop. However, notably towards the catchment divide on the north part of the map, the topographic index has a complex pattern, associated with small areas of both low and high values of the topographic index, and leading to the appearance of disconnected saturated areas shown in Figure 2a and 2b.

Figure 4 shows a series of zero/negative saturation deficit predictions, for the same time period as shown in Figure 2b, in which the topographic data are re-mapped at progressively coarser resolutions by spatial averaging of elevations within each cell. This shows that as the topographic resolution is coarsened, the number and extent of unconnected saturated areas are reduced: the catchments display more coherent patterns, with saturated areas more effectively connected to the channel network. This is not new: Quinn *et al.* (1997) showed how progressively fining model resolution from 50 m to 5 m reduces the kurtosis in the distribution of topographic index values and increases quite substantially the number of very low index values. It was noted that this was because fining model resolution tends to increase the number of cells with a low value of the upslope contributing area whilst leaving the overall slope gradient the same. Gullies and cells on or close to the channel network will be less affected by this than other parts of the hillslope. Thus, coarser grid cells are biased to higher values of the topographic index and result in a larger saturated area for a given storm. Such cells also more commonly form coherent patterns of zero saturation deficit clustered on the channel network (*cf.* Figure 4).

As the resolution of topographic data is fined, so there is a greater possibility that the DEM includes areas of natural depression storage in terms of both surface flow and saturated zone flow. The requirement here is to separate areas of natural depression storage from artificial depressions that result from random errors in the topography defined by the estimated elevation precision. Lane *et al.* (2003) show that under the assumption that the global uncertainty in surface quality (as represented by the standard deviation of error in elevation, $\sigma_e$) applies to individual elevations, then the uncertainty in slope calculated between any two adjacent elevations ($\sigma_s$) has a grid spacing ($d$) dependence:

$$\sigma_s = \frac{\sqrt{2}\sigma_e}{d}$$

Thus, as grids are fined, there is also a greater probability of error in slope even if there is no change in data point quality.

Figure 5 illustrates a simple experiment to assess this. It is based on a DEM with 16 x 16 pixels and a 2.0 m spacing describing a north-south trending 5.7° slope with a small headcut. A random error was then added to each elevation, drawn from a normal distribution defined by the precision of the LIDAR DEM used in this study (±0.12 m). The number of pits that resulted was counted and this was repeated 2,500 times to determine the average percentage of pits associated with this random elevation error. This was 8.8 % or approximately 23 grid cells. The experiment was then repeated on a DEM smoothed to a 4.0 m spacing and, on average, 0.8 % or 0.54 grid cells were found to be pits. The extent to which pits are found has a slope dependence, with a higher percentage of pits in areas of lower slope. Unfortunately, areas of lower slope are also more likely to contain zones of depression storage. Thus, as the resolution of the topographic data becomes finer, the traditional approach to sink removal becomes problematic as the distinction between artificial and real depression storage zones is blurred. Alternative methods for dealing with sinks are therefore required.

Even ignoring the issue of artificial sinks and the effects of topographic data response upon the dynamics of saturated areas, Figure 2 suggests that to assume that all areas with a zero or negative saturation deficit are connected to the channel network is not necessarily correct. This will lead to certain areas contributing runoff too quickly. It will also change the rate at which the catchment becomes saturated (following from [1]) as unconnected saturated areas are assumed to be saturated and to contribute to overland flow when water might otherwise re-infiltrate in zones of lower topographic index before the channel is reached. The original TOPMODEL representation is likely to be of particular concern in two situations:

(1) where land drainage effects are to be simulated . A major effect of land drainage is upon the extent to which areas prone to high saturation connect with the drainage network (Holden *et al.*, in press). Despite the presence of open drains, subsurface hydrology may still be strongly topographically controlled where lateral saturated permeability is low (Holden and Burt, 2003a; 2003b). Whilst parameterisation may be used to represent the effects of processes like land drainage upon output hydrographs, we wish to investigate hydrograph response with and without land drainage, so that the effects upon connectivity need to be explicitly represented; and

(2) where there is a particular interest in water quality or sediment delivery problems, and identification of source areas is of particular importance. To model the transfer of faecal coliforms and other organisms on the soil surface, it is also necessary to have an explicit representation of how saturated areas are or are not connected to streams.

To meet these requirements, a network index treatment is developed in the next section that allows for the dynamic expansion of connected saturated areas to the drainage network. We also identify a method for dealing with the problem of pits in the DEM.

**Development of a network index approach for the treatment of disconnected saturated areas**

The basic principle of the network index approach is straightforward: a saturated area can only connect to the drainage network when all cells in the model between the saturated area and the network are themselves saturated. We begin by making a simple assumption: that each grid cell in the model has a single flow path that can be defined using the line of steepest descent. Support for this assumption is provided by Quinn *et al* (1991, 1997). Quinn *et al* (1991) developed a multiple flow routing algorithm, which shared upslope contributing area to grid cells on the basis of the shape of the contours that pass through a grid cell in relation to its neighbours. Quinn *et al* (1997) show that the Holmgren (1994) function can be used to generalise this algorithm from multiple algorithms through to single flow direction algorithms:

$$d_{ik} = c_i \tan \beta^h \qquad [2]$$

where; $d_{ik}$ is the flow proportion for cell $i$ in the $k$th downslope direction; $c_i$ is the contour length for cell $i$; and $h$ is a parameter. For $h = 1$, [2] gives the original multiple flow algorithm of Quinn *et al* (1991); for $h = 100$, it gives a single flow algorithm based upon routing all water along the line of steepest slope. Quinn *et al.* (1997) found that with a 5 m DEM, a value of $h = 50$, which is almost a single flow algorithm, gave a reasonable reproduction of the spatial patterns of topographic index as compared with the multiple flow algorithm case. They concluded that, for larger pixels, lower values of $h$ will be required in order to get an effective representation of the route taken by water over the hillslope surface. Although a single flow algorithm still provides a coarse discretization of flow, as smaller areas are being routed, the cumulative error will be reduced. In this application, the 2.0 m resolution makes it increasingly probable that a single flow algorithm based upon steepest descent will give an adequate representation of routing. By using a minimum upslope contributing area, the steepest flow direction map provides a first approximation of a channel network. This is defined from $\tan^{-1}(\Delta z/\Delta h)$ where $\Delta z$ is the elevation difference and $\Delta h$ is the horizontal difference. For the four orthogonal neighbours $\Delta h$ is the cell size. For the four diagonal neighbours $\Delta h$ is square root 2 times the cell size.

Flow direction is used to define internally draining areas or hollows. There are three issues with getting flow paths from the DEM: pits, flats and hollows. Pits are single cells that are local elevation minima. Flats are adjacent cells with equal elevation and are a problem if a cell and all its neighbours have equal elevation because there is no flow direction from the central cell. Hollows are contiguous cells that have no external drainage i.e. all the flow directions are internal. These cases are dealt with in three steps. First, we fill pits by setting the elevation equal to the lowest neighbour. This creates a flat and there may be other flats in the original data. Second, we deal with flats. This is based upon finding the exit associated with a given flat area and then assigning flow directions towards that exit. The process uses two auxiliary raster layers. The first is flow direction and the second is a binary layer with flat cells marked with a minus one and other cells a zero. For each marked cell on the flats layer, we check if there is a non-zero value on the flow direction layer. If there is, we mark the cell with a 1 on the flats layer. Next, we mark all flat neighbours of the 1 marked cells with a 2. This is repeated for the 2's setting neighbours to 3 and so on until all the –1 cells on the flats layer have all been re-marked. Finally, again using the auxiliary raster, we define a flow direction for each cell marked 2 or greater to its lowest valued neighbour. The 1 marked cells already have a flow direction so every cell now has a flow direction. Third, there may still be internal drainage areas or hollows. These are dealt with as follows. We define a sink raster that defines where the cells must drain to for each sub-catchment. This scan be made from a channel map. We then mark the catchment by tracing flow directions upslope from the sink cells. Cells in hollows will be unmarked because the flow directions do not lead to the sink. Each hollow is now examined to find the lowest cell on its boundary. If this cell is adjacent to a catchment cell the hollow is incorporated into the catchment by tracing back to the centre of the hollow, reversing flow directions, and marking cells along the way. This process is repeated until no more hollows can be added to the catchment.

The network index approach then proceeds in a straightforward manner. Starting at each channel cell, each sub-catchment network is traced upslope, recording the lowest value of the topographic index encountered along this flow path. This lowest value is assigned to each new cell encountered upstream along the follow path until a lower value is encountered. This can be done because of the use of a steepest direction flow algorithm, which assumes that each cell has a unique downslope direction, which is visited only once. Thus, each grid cell has both a topographic index value, that can be used to identify when the saturation deficit is zero or negative and overland flow can occur, and a network index value, which defines when a cell with a zero or negative saturation deficit is connected to the drainage network. There is an important link here back into the moisture accounting embodied in [1]. In process terms, we assume that if the local saturation deficit is zero or negative, but an area of overland flow is not network connected, then overland flow will continue to the point along the flow path that has a positive saturation deficit where infiltration will occur. This is represented in [1] by assuming that any water that is predicted to be unconnected overland flow is assumed to contribute negatively to the average catchment storage deficit. Effectively, this reduces the critical topographic index above which locations within the catchment have a zero or negative storage deficit: i.e. it makes those locations with a higher topographic index saturate sooner. This assumption is most likely to be acceptable where there is a smooth variation of the topographic index in space, and a given flow path is not punctuated by an exceptionally low value of the topographic index. If it is, high rates of overland flow delivery should lead to rapid reduction in the local saturation deficit at that point such that its topographic index is not a good indicator of its propensity to saturate. We deal with this problem by making sure that there is sufficient sub-division of the catchment over which the lumping (i.e. [1]) is applied. Ultimately, this treatment can be viewed as one end of a spectrum of two extremes. In the basic form of TOPMODEL, all runoff that is generated by locations with zero or negative saturation deficit becomes overland flow. In the network index based approach, runoff only becomes overland flow if all cells along the flow path to the channel network can also generate overland flow. If not, it is assumed to go into storage within the catchment. When each sub-catchment's average storage deficit is high, the network index approach is most likely to be effective as the probability of re-infiltration of overland flow along a flow path is highest. As the average storage deficit falls, so there is a greater probability that the network index over-estimates the re-infiltration of flow.

For the storm event in Figure 3, Figure 6 shows a series of predictions through time of unsaturated, unconnected saturated and connected saturated areas. For the particular combination of parameter values used ($m$ = 0.003 m; $T_o$ = 1.0 $m^2s^{-1}$) this shows that the spatial extent of saturation increases rapidly. However, the spatial extent of connected saturated zones increases to a much lesser extent. It appears that proximity to the channel network is a crucial control upon connectivity because it reduces the probability that there is a zone of low topographic index values that prohibits connection. Connectivity tends to be strongly associated with the drainage network in relation to local topographic slope. Close proximity to the drainage network increases the probability that upslope contributing area is sufficient to raise the topographic index. Whether or not this is the case is determined by the local topographic slope, which may be sufficient to reduce the topographic index to lower values and hence prevent surface connectivity (even if subsurface drainage rates are then quite high). This implies very strong sensitivity to the topographic detail in the model and implies that close inspection of data quality, coupled to thoughtful parameterisation, will be required as this model is properly tested. The rate of response shown in Figure 6 is rapid in terms of the generation of both saturated areas and connected saturated areas. This would be expected given the formulation of the network index model. Connection and disconnection should be a rapid process as it is driven by the catchment averaged saturation deficit in relation to the critical index value along a given flow path. Small changes in saturation deficit will lead to quite considerable changes in the extent of connectivity as low values of the topographic index along a given flow path may affect large upslope contributing areas. Indeed, the connectivity declines more rapidly than saturation. The question is whether or not this is leading to a better representation of hydrological response. It should be noted that the results in Figure 6 involve a three-fold classification (unsaturated, saturated unconnected, saturated connected) and the Figure implies that extensive saturation and, in particular, extensive connected saturation is a short lived phenomenon, dependent upon maintenance of continued precipitation. This matches preliminary field results from a distributed network of water level recorders that show that extensive areas of negative saturation deficit are short-lived except where the topographic index is sufficiently high to allow significant flow accumulation (e.g. at the foot of hillslopes). One issue is currently under investigation. There will be strong (and potentially nonlinear) feedback from the network index through [1] where unconnected saturated areas reduce saturation deficit. As the base flow component has a negative exponential dependenct upon saturation deficit, this may rapidly increase the magnitude of the baseflow component. This may artificially increase the rate of baseflow generation meaning that additional attention needs to be given to the definition of individual sub-catchments and, possibly, the isochrone treatment of baseflow reported by Saulnier *et al.* (1997).

**Implications**

The development of the network index approach has a number of methodological and substantive implications. In methodological terms, it will be important to undertake a GLUE-type analysis to explore the nature of parameter sensitivity with the network-based model. This is for three reasons. First, whilst there is no increase in the number of parameters that the model requires, the hydrological response will be very sensitive to the minimum value of the topographic index encountered along a given flow path, and hence parameters that control the topographic index. Second, and related to this, the network index approach gives greater emphasis to a relatively small number of cells, which exert a critical control upon the connectivity of saturated areas. This was illustrated in Figure 6, where often only a small number of grid cells with low values of the topographic index can prevent quite large saturated areas from connecting to the drainage network. The use of high resolution data diminishes this problem somewhat, but it is of increased importance to propagate topographic uncertainty in order to account for the effects of this upon those cells that exert a major control upon connectivity.

In substantive terms, the model provides a tool for explicit representation of scenarios where land management processes involve changes in the way in which saturated zones connect to the drainage network. For the parameter values used in Figure 6, it is clear that even during a storm event, there may be large areas of the landscape that are not saturated or, more importantly, saturated but not connected. Until a full parameterisation exercise has been undertaken, it will not be possible to assess the extent to which this is a possible system characteristic. However, work by Holden and Burt (2003a) and Holden and Burt (in press) in upland blanket peats suggests that it is, which may have major implications for our current conception of what causes flood events. Even if the method is shown to be of relatively poor importance for understanding flood generation processes, it will allow a more correct representation of the partitioning of surface and subsurface flow, and especially of situations where overland flow generated in one part of the catchment becomes surface flow as it moves to the drainage network. In doing so, it will acquire chemical and biological signatures that may make the network index representation of saturated unconnected areas especially important.

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**List of Figures**

Figure 1. A shaded relief map of the 2.0 m DEM used in this study.

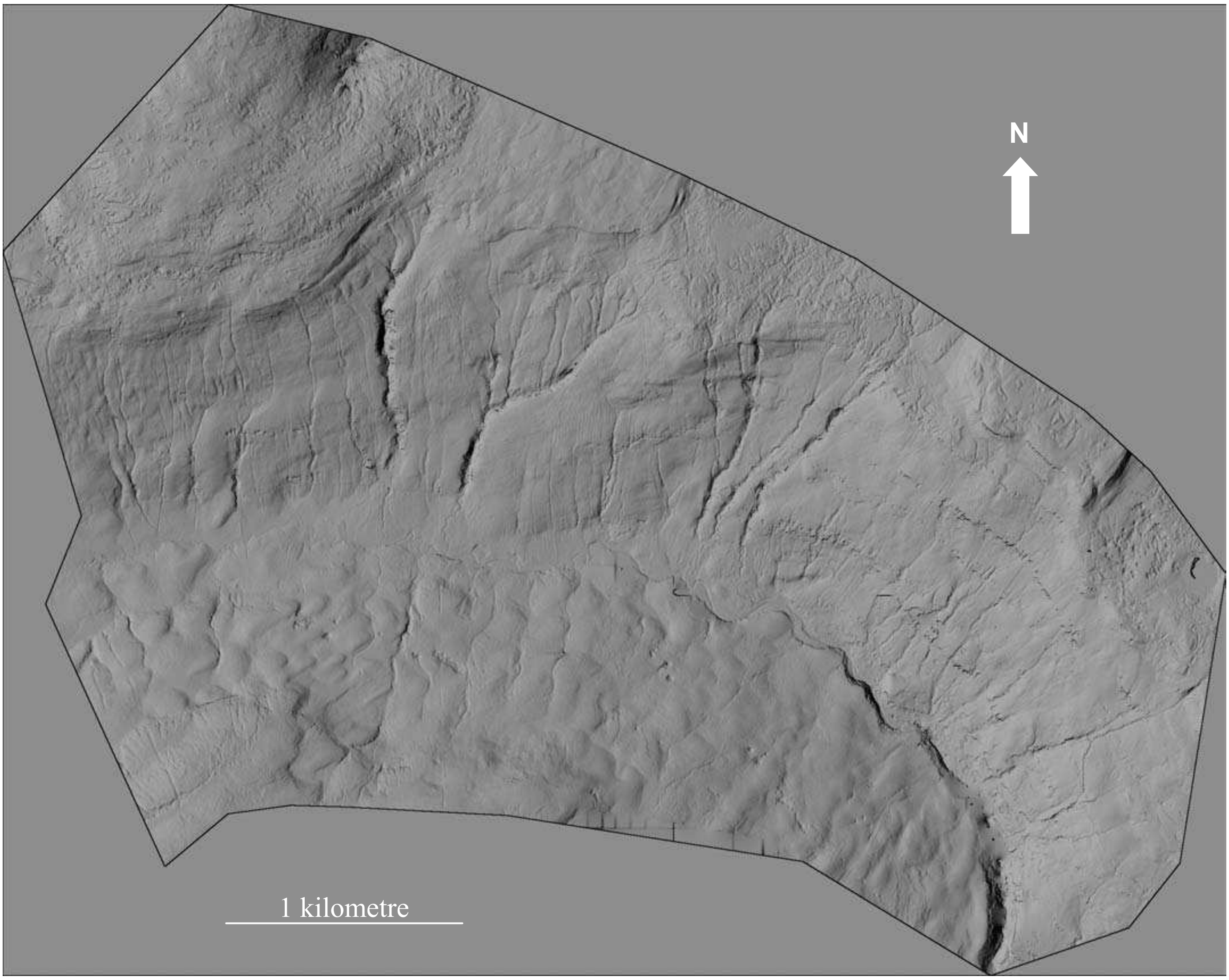
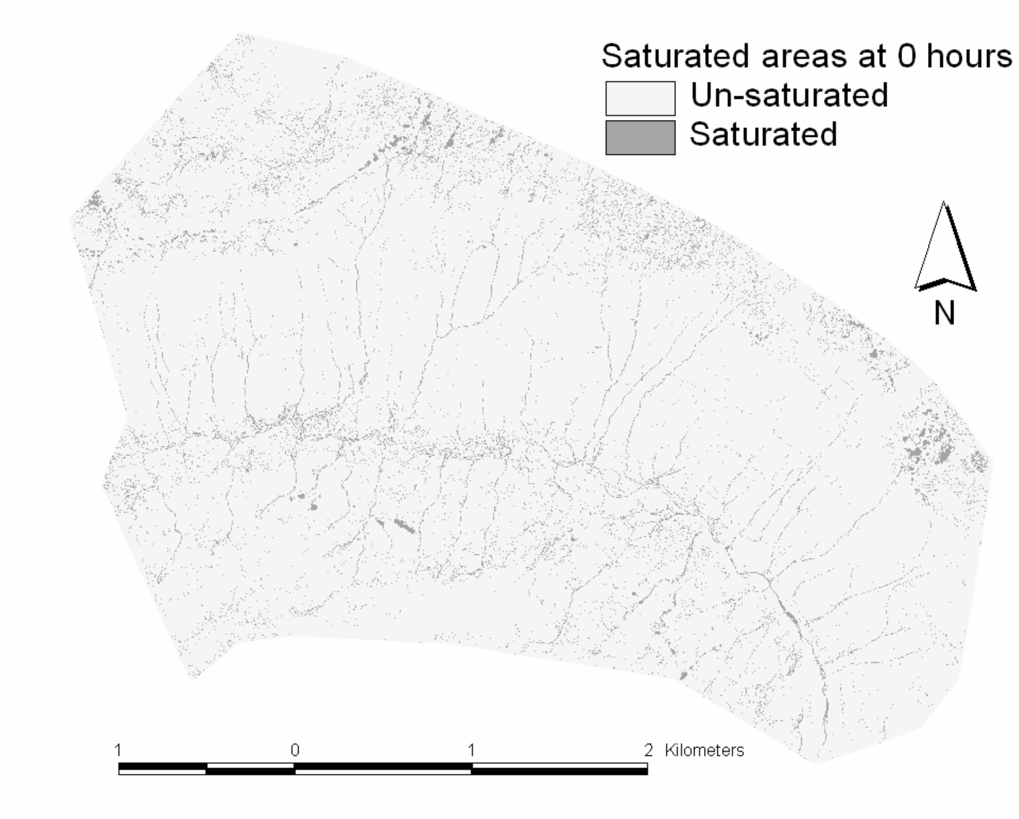
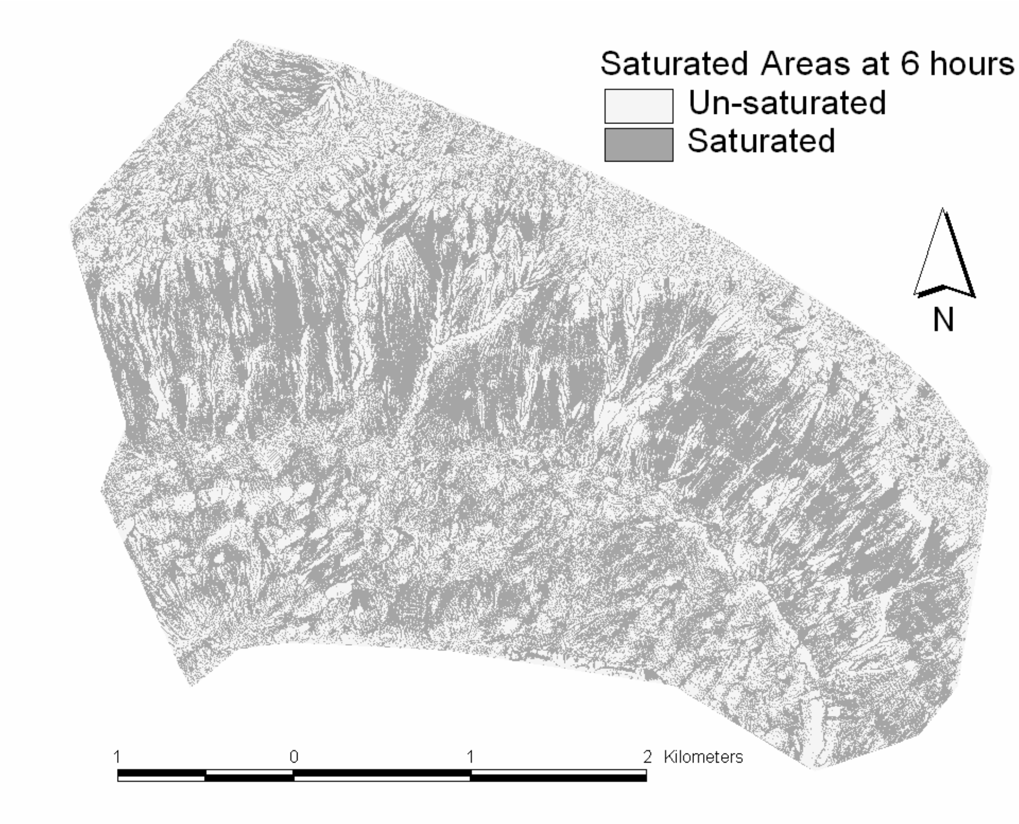


Figure 2. Predictions of zero saturation deficit at the start (Figure 2a) and after 6 hours after onset of rain (Figure 2b) for a storm shown in Figure 3. Figure 2c shows the associated topographic index.



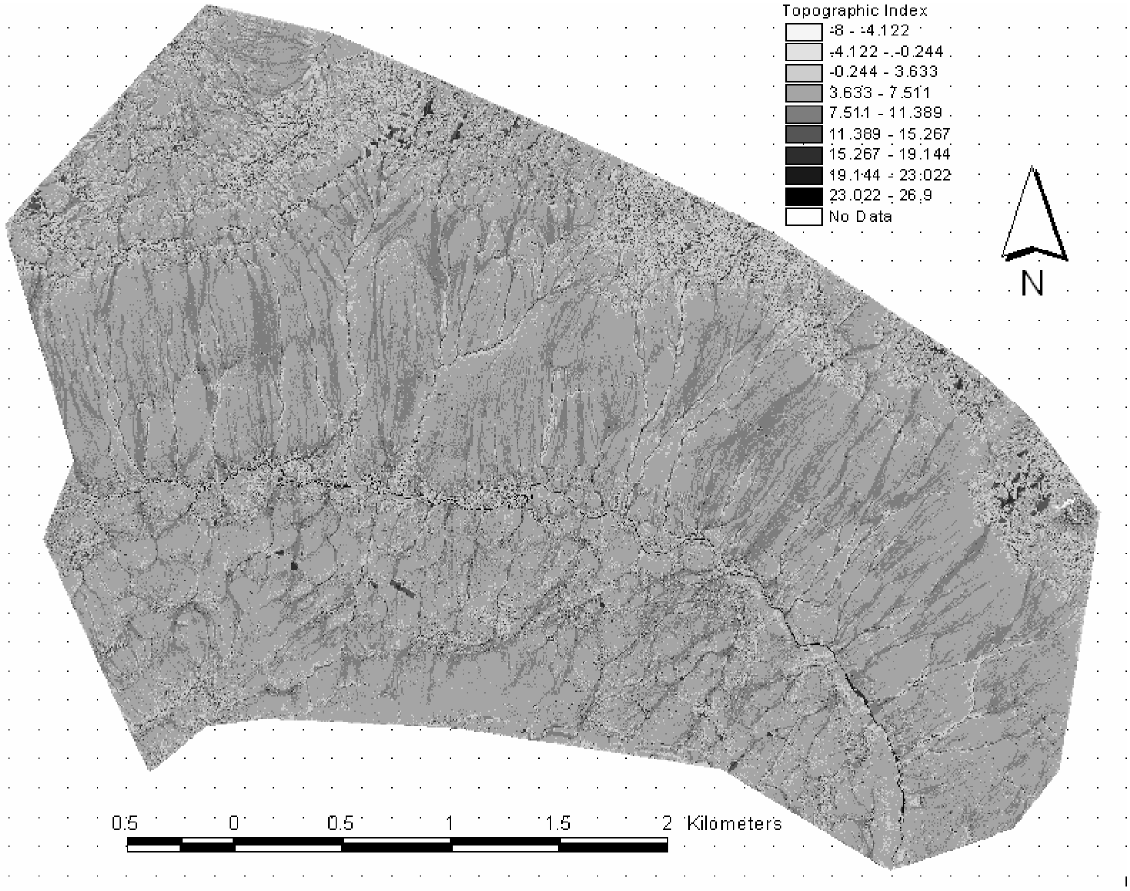

Topographic Index
-8 - -4.122
-4.122 - -0.244
-0.244 - 3.633
3.633 - 7.511
7.511 - 11.389
11.389 - 15.267
15.267 - 19.144
19.144 - 23.022
23.022 - 26.9
No Data
N
0.5 0 0.5 1 1.5 2 Kilometers

Figure 3. The June 2000 storm event used to generate the predictions of zero/negative saturation deficit shown in Figure 2. Increments are 15 minutes.

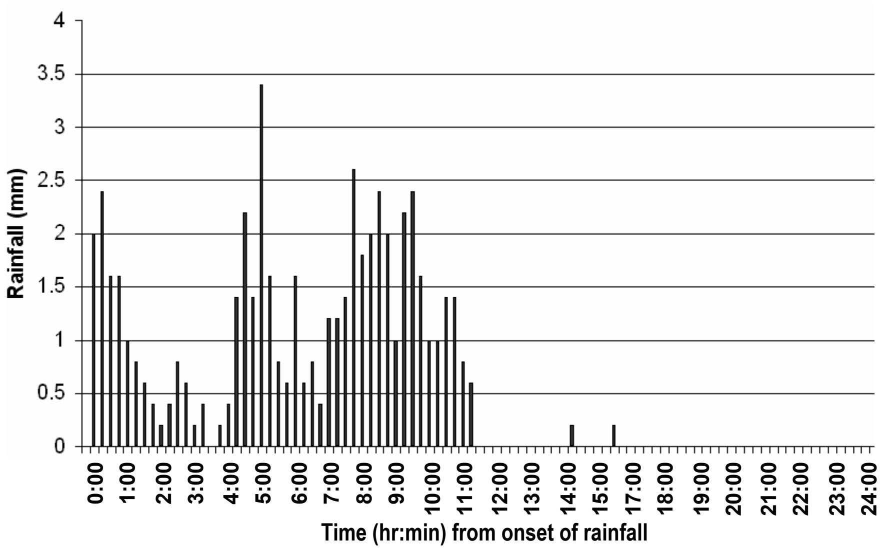


Figure 4. The effects of DEM resolution upon patterns of saturation for 2.0 m, 4.0 m, 8.0 m, 16.0 m, 32.0 m and 64.0 m DEMs.

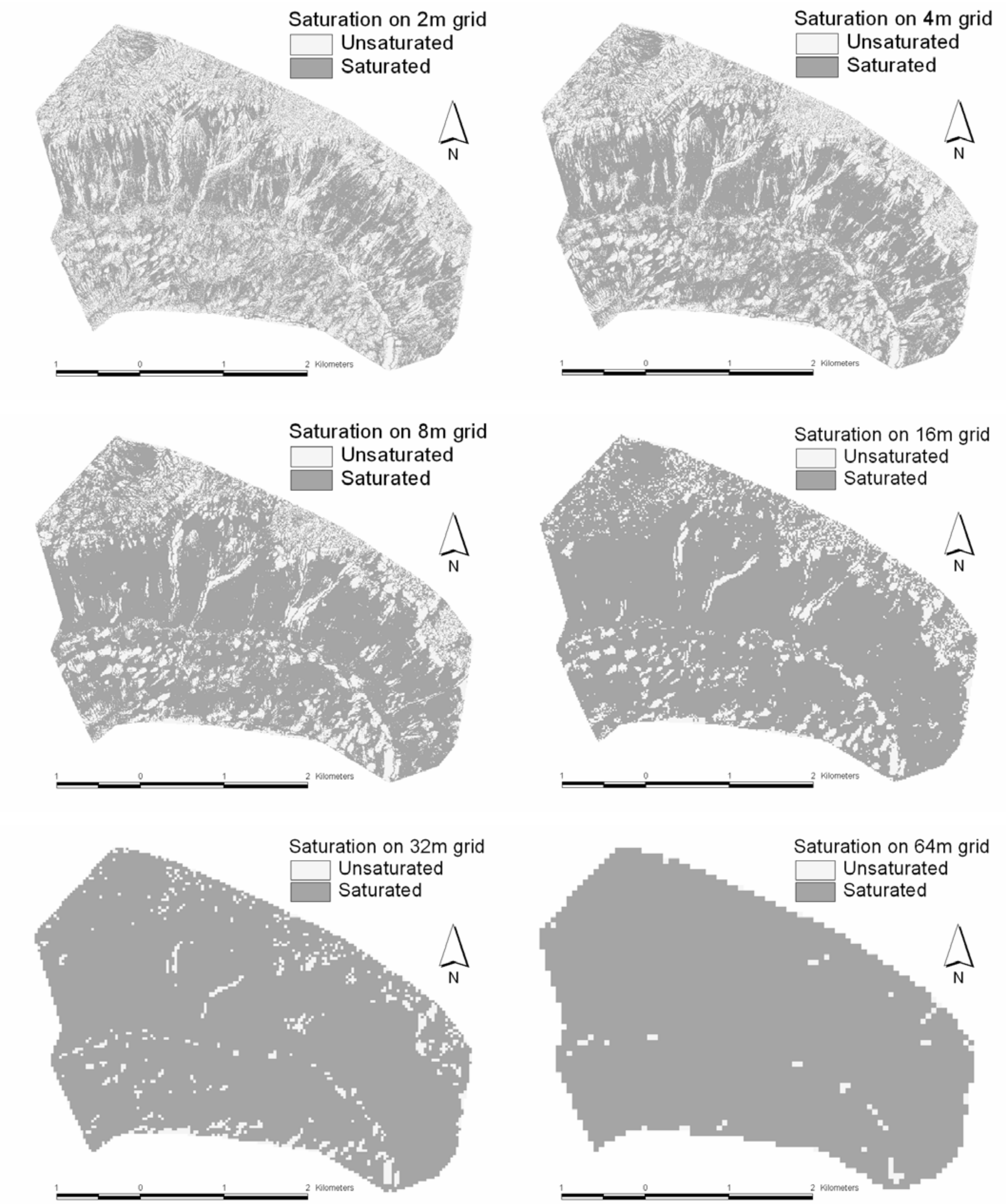


Figure 5.  The DEM used to illustrate the resolution dependence of pit occurrence (Figure 5a) and one example of a perturbed DEM (Figure 5b).

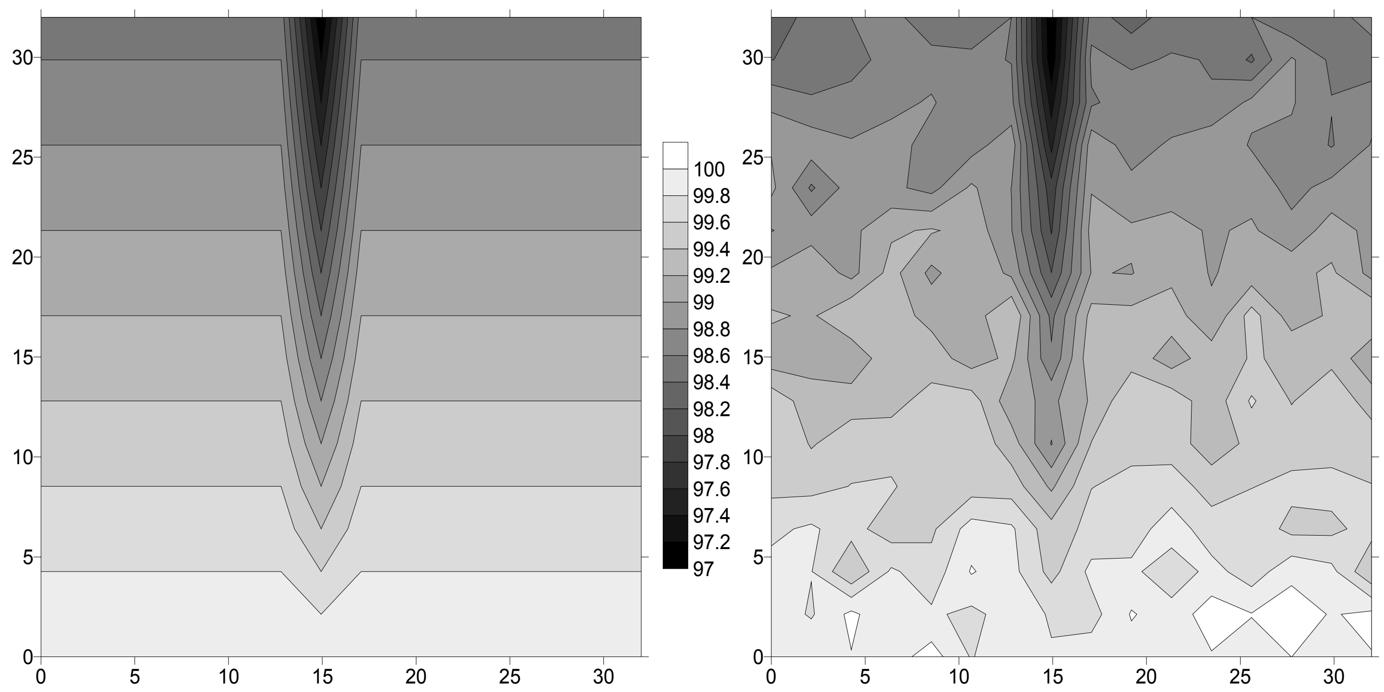
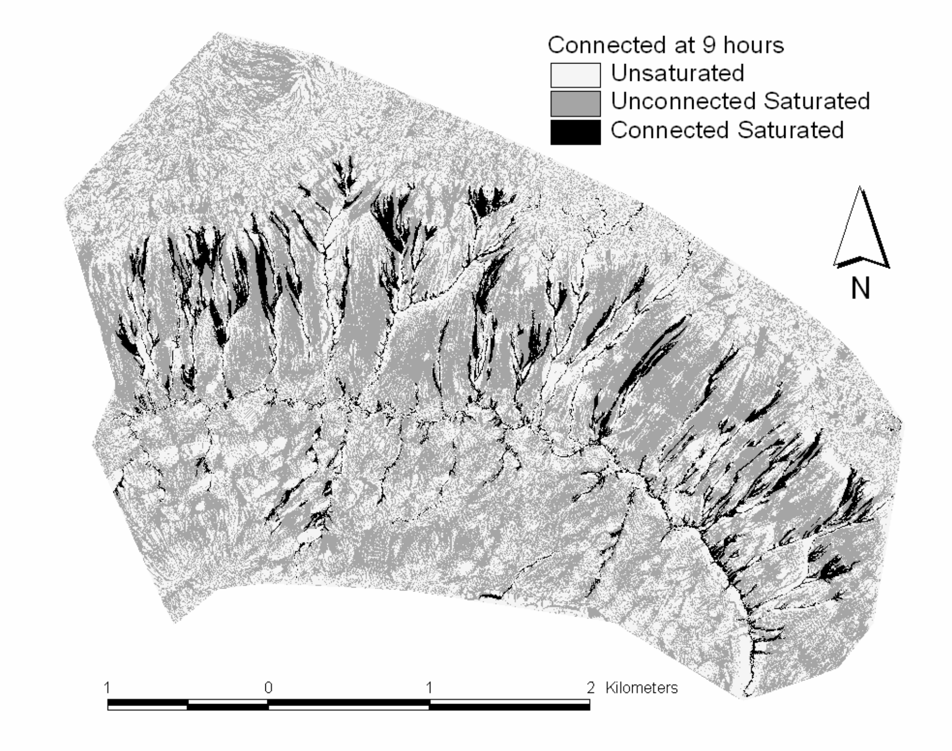
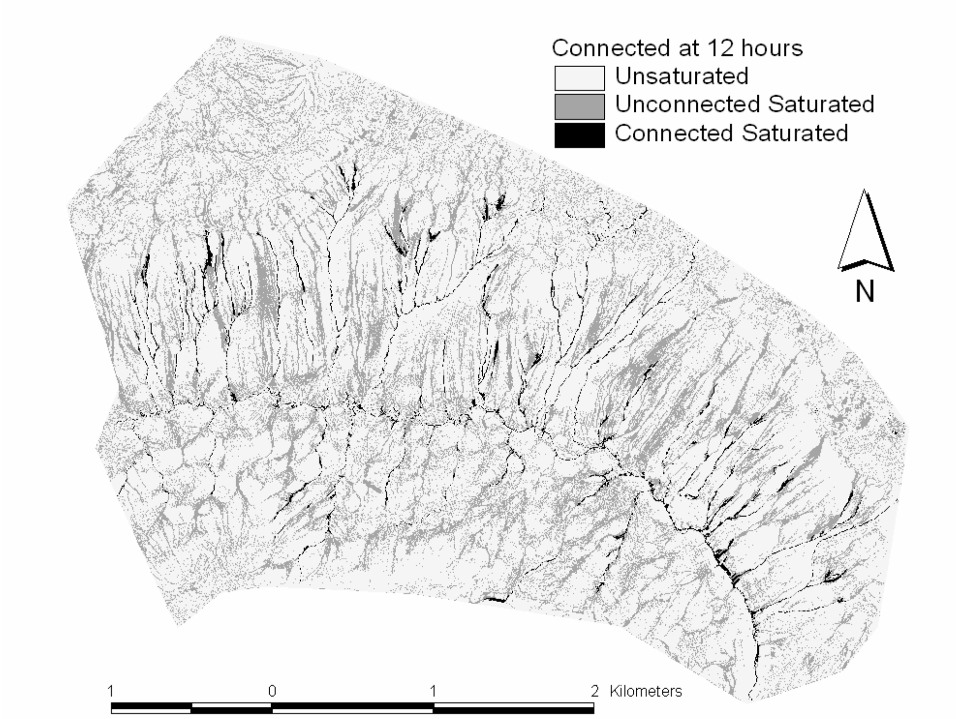
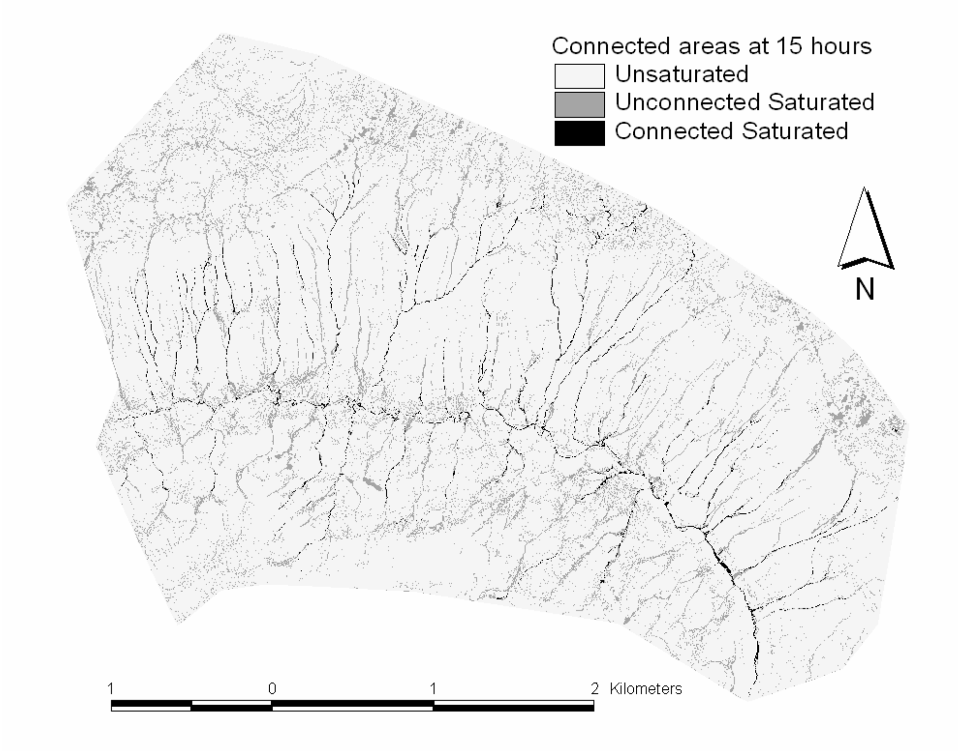


Figure 6. Predictions through time of unsaturated, unconnected saturated and connected saturated areas for 6, 9, 12 and 15 hours from the onset of rainfall for the June 2000 storm event shown in Figure 3.


Connected at 9 hours
Unsaturated
Unconnected Saturated
Connected Saturated
N
1 0 1 2 Kilometers


Connected at 12 hours
Unsaturated
Unconnected Saturated
Connected Saturated
N
1 0 1 2 Kilometers


Connected areas at 15 hours
Unsaturated
Unconnected Saturated
Connected Saturated
N
1
0
1
2 Kilometers