# Terrain-Based Descriptions of Earth Surface and Atmospheric Processes

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**Abstract:** Terrain analysis offers a robust method for modelling over large areas the complex spatial patterns of environmental systems. This paper provides a guide to the use of terrain analysis methods for the study of spatial patterns and processes in a range of landscapes and application domains. It illustrates some of the applications of terrain analysis, explains the algorithms used by current terrain analysis software, and deals with details such as the quirks of particular algorithms, interpretation of terrain attributes, use of terrain attributes in predictive models, and the effects of scale and resolution. This paper describes the state of the art in DEM generation, terrain analysis methods, and applications.

## 1. Introduction

Terrain plays a fundamental role in modulating earth surface and atmospheric processes (Wilson and Gallant 2000). This linkage is so strong that an understanding of the nature of terrain can directly confer understanding of the nature of these processes, in both subjective and analytical terms, at numerous spatial and temporal scales (Hutchinson and Gallant 2000).

Most of the hydrologic, geomorphic, and ecological research of the past century has been conducted at the global and nano- or micro-scales identified in Figure 1. The meso- and topo-scales have received much less attention, and yet these scales are important because many of the solutions to environmental problems such as accelerated soil erosion and non-point source pollution will require changes in management strategies at these landscape scales (Wilson et al. 2000).

The influence of geologic substrate on soil chemistry (e.g. Likens et al. 1977) and impact of prevailing weather systems and elevation-driven lapse rates on long-term average monthly climate (e.g. Hutchinson 1995) exemplify some of the controls operating at the meso-scale. The influence of surface morphology on catchment hydrology and impact of slope, aspect, and horizon shading on insolation probably represent the most important controls operating at topo-scales. Numerous studies have shown how the shape of the land surface can affect the lateral migration and accumulation of water, sediments, and other constituents (e.g. Moore et al. 1988). These variables, in turn, may influence soil development (e.g. Kreznor et al. 1989) and exert a strong influence on the spatial and temporal distributions of the light, heat, water, and mineral nutrients required by photosynthesizing plants (e.g. Franklin 1995; Mackey 1996).

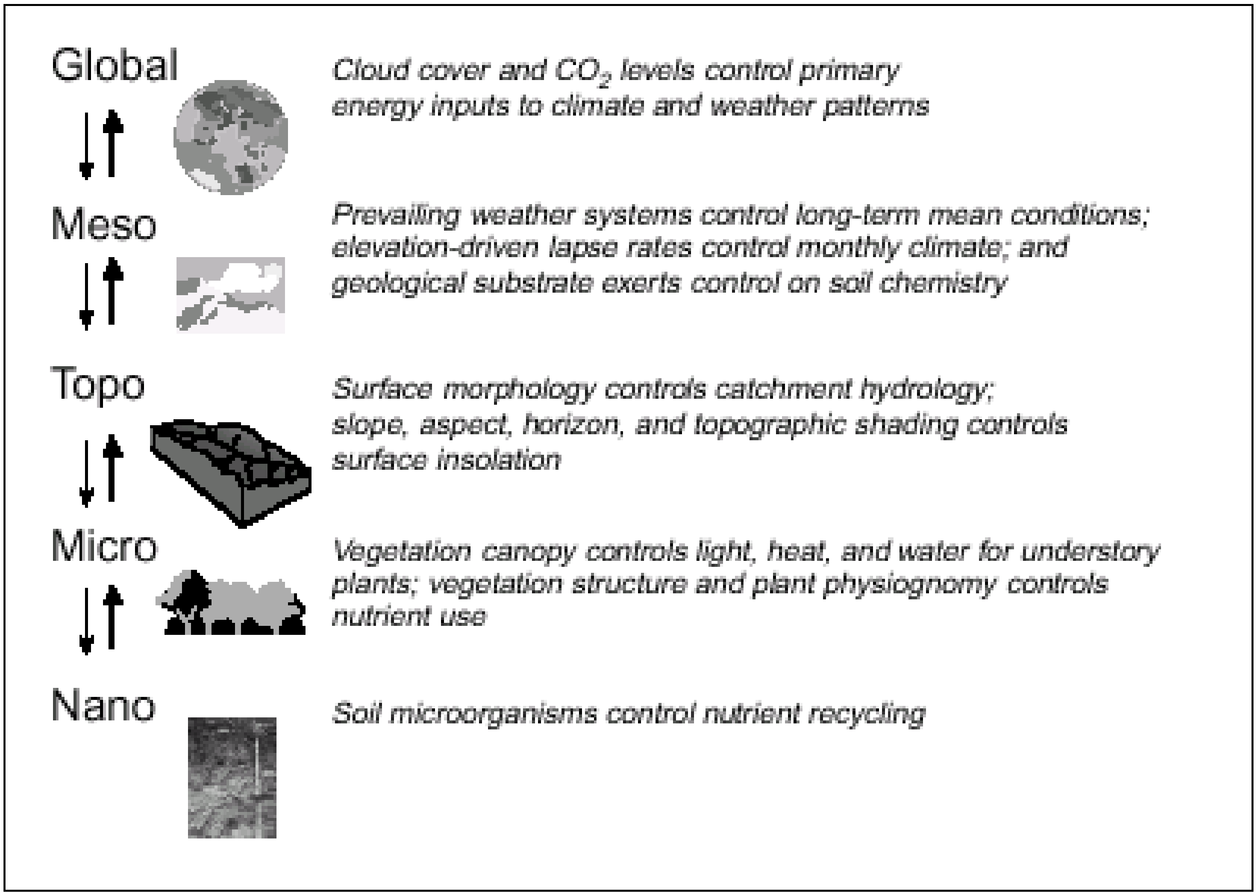


**Figure 1.** The scales at which various biophysical processes dominate calculation of primary environmental regimes (from Mackey 1996)

The increased popularity of work at these two intermediate scales during the past decade has capitalized on the increasing availability of high resolution, continuous, digital elevation data and development of new computerized terrain analysis tools (Wilson 1996; Burrough and McDonnell 1998). This paper aims to describe these new data sources, the terrain attributes that can be calculated from them, and some of the problems and shortcomings that will have to be solved for these data to be used to describe several key environmental processes and patterns accurately in the future.

## 2. Digital Elevation Data Sources and Structures

There are three main classes of source topographic data (Hutchinson and Gallant 2000). The first class consists of surface specific point elevations and may be obtained by ground survey and photogrammetric data capture. These point data are an ideal data source for most interpolation techniques, including triangulation methods and specially adapted gridding methods. The advent of the global positioning system (GPS) has enhanced the availability of accurate ground surveyed data for small experimental catchments but they are seldom used for large areas (Fix and Burt 1995, Twigg 1998, Wilson 1999). The second class consists of contour data. These are still the most common terrain data source for larger areas and the conversion of contour maps to digital form is a major activity of mapping organizations worldwide (Hobbs 1995). Contours can also be generated automatically from photogrammetric stereo models, although these methods may be subject to error due to variations in surface cover and the paucity of contour lines in areas of low relief.

The final class is gridded DEMs that are calculated directly from stereoscopic interpretation of data collected by airborne and satellite sensors. The traditional source of these data is aerial photography, although the same stereoscopic methods have been applied more recently to SPOT imagery and both airborne and spaceborne synthetic aperture radar (SAR) (Hutchinson and Gallant 2000). Airborne and light detection and ranging (LIDAR) scanners provide another promising data source that may help with both the generation of "bare" surface DEMs and the identification and classification of surface features (e.g. vegetation, buildings) (Hill et al. 2000). All remote sensing methods measure elevations with significant random errors, which depend on the inherent limitations of the observing instruments, as well as surface slope and roughness (Harding et al. 1994, Dixon 1995) and careful filtering and interpolation of such data will usually be required to derive useful representations of surface shape and drainage structure (Hutchinson and Gallant 2000).

Their widespread availability, simplicity (i.e. simple elevation matrices that record topological relations between data points implicitly) and ease of computer implementation help to explain why square-grid DEMs have emerged as the most widely used data structure during the past decade (Moore et al. 1991, 1993e; Wise 1998). These advantages are offset by at least three disadvantages. First, the size of the grid mesh will often affect the storage requirements, computational efficiency, and quality of the results (Collins and Moon 1981; Moore et al. 1991). Second, square grids cannot handle abrupt changes in elevation easily and they will often skip important details of the land surface in flat areas (Carter 1988). However, it is worth noting that many of the problems in flat areas occur because the USGS and others persist in recording elevations in whole meters. Third, the computed upslope flow paths will tend to zigzag across the landscape and increase the difficulty of calculating specific catchment areas accurately (Moore et al. 1991). Several of these obstacles have been overcome in recent years: for example, there is no generic reason why regular DEMs cannot represent shape well in flat areas so long as the terrain attributes are calculated by a method that respects surface drainage. Similarly, the advent of several new compression techniques have reduced the storage requirements and improved computational efficiency in recent years (e.g. Smith and Lewis 1994). DEMs with grid sizes of 500 m, 100 m, 30 m, 10 m, and even 1 m are increasingly available for different parts of the globe (see U.S. Geological Survey 1993, Ordinance Survey 1993, Hutchinson et al. 1996 for examples).

Triangulated Irregular Networks (TINs) have also found widespread use (e.g. Tajchman 1981, Jones et al. 1990, Yu et al. 1997; Kidner et al. 2000). TINs are based on triangular elements (facets) with vertices at the sample points (Moore et al. 1991). These facets consist of planes joining the three adjacent points in the network and are usually constructed using Delauney triangulation (Weibel and Heller 1991). Lee (1991) compared several methods for building TINs from gridded DEMs. However, the best TINs sample surface-specific points such as peaks, ridges, and breaks in slope and form an irregular network of points stored as a set of x, y and z values together with pointers to their neighbors in the net (Moore et al. 1991). TINs can easily incorporate discontinuities and may constitute efficient data structures because the density of the triangles can be varied to match the roughness of the terrain (Moore et al. 1991). This arrangement may cancel out the additional storage that is incurred when the topological relations are computed and recorded explicitly (Kumler 1994).

The proliferation of digital elevation sources and preprocessing tools means that the initial choice of data structure is not as critical as it once was. Numerous methods have been proposed to convert digital elevation data from one structure to another although care must be exercised with each of these methods to minimize unwanted artifacts. In addition, larger quantities of data do not necessarily produce better results: Eklundh and Martensson (1995), for example, used ANUDEM (Hutchinson 1989) to derive square grids from contours and demonstrated that point sampling produces faster and more accurate square-grid DEMs than the digitizing of contours. Similarly, Wilson et al. (1998) used ANUDEM to derive square grids from irregular point samples and showed that many of the x, y and z data points acquired with a truck-mounted GPS were not required to produce satisfactory square-grid DEMs. ANUDEM is one of many interpolation methods that has been proposed and is unique in the way it calculates ridge and streamlines from points of maximum local curvature on contour lines and incorporates a drainage enforcement algorithm that automatically removes spurious sinks or pits in the fitted elevation surface.

## 3. Primary Topographic Attributes

We usually distinguish primary attributes that are computed directly from the DEM and secondary or compound attributes that involve combinations of primary attributes and constitute physically-based or empirically derived indices that can characterize the spatial variability of specific processes occurring in the landscape (see Wilson and Gallant 2000 for additional details). This same logic is adopted here.

The primary attributes include slope, aspect, plan and profile curvature, flow path length, and upslope contributing area (see Table 1 for a more complete list). Most of these topographic attributes can be derived from all three types of elevation data for each and every element (i.e. grid cell or triangular facet) as a function of its surroundings (Moore et al. 1991, 1993e). Although the methods for calculating these attributes are well known, some decisions related to the removal of depressions from the DEMs and the selection of one or more rules to determine drainage directions, the connectivity of individual elements, and calculation of flow path lengths and upslope contributing areas will usually be required (e.g. Quinn et al. 1991; Costa-Cabral and Burges 1994; Tarboton 1997). The available flow routing methods will produce large variations in flow direction and upslope contributing area grids, although it is not clear which method is most appropriate and/or accurate in specific landscapes (see Wolock and McCabe 1995, Desmet and Govers 1996a and Wilson et al. 2000 for comparisons of existing flow routing algorithms).

**Table 1.** Primary topographic attributes calculated from DEM data (from Wilson and Gallant 2000)

| Attribute | Definition | Significance |
|---|---|---|
| Altitude | Elevation | Climate, vegetation, potential energy |
| Upslope height | Mean height of upslope area | Potential energy |
| Aspect | Slope azimuth | Solar insolation, evapotranspiration, flora and fauna distribution and abundance |
| Slope | Gradient | Overland and subsurface flow velocity and runoff rate, precipitation, vegetation, geomorphology, soil water content, land capability class |
| Upslope slope | Mean slope of upslope area | Runoff velocity |
| Dispersal slope | Mean slope of dispersal area | Rate of soil drainage |
| Catchment slope | Average slope over the catchment | Time of concentration |
| Upslope area | Catchment area above a short length of contour | Runoff volume, steady-state runoff rate |
| Dispersal area | Area downslope from a short length of catchment | Soil drainage rate |
| Catchment area | Area draining to catchment outlet | Runoff volume |
| Specific catchment area | Upslope area per unit width of contour | Runoff volume, steady-state runoff rate, soil characteristics, soil water area content, geomorphology |
| Flowpath length | Maximum distance of water flow to a point in the catchment | Erosion rates, sediment yield, time of concentration |
| Upslope length | Mean length of flow paths to a point in the catchment | Flow acceleration, erosion rates |
| Dispersal length | Distance from a point in the catchment to the outlet | Impedance of soil drainage |
| Catchment length | Distance from highest point to outlet | Overland flow attenuation |
| Profile curvature | Slope profile curvature | Flow acceleration, erosion/deposition rate, geomorphology |
| Plan curvature | Contour curvature | Converging/diverging flow, soil water content, soil characteristics |
| Tangential curvature | Plan curvature multiplied by slope | Provides alternative measure of local flow convergence and divergence |
| Local topographic position | Proportion of cells in a user-defined circle lower than the center cell | Relative landscape position, flora and fauna distribution and abundance |

The overall aim is to be able to use the computed attributes to describe the morphometry, catchment position, and surface attributes of hillslopes and stream channels comprising drainage basins (e.g. Band 1993a, b; Jenson and Domingue 1988; Montgomery and Foufoula-Georgiou 1993). Some authors, such as Dikau (1989), Dymond et al. (1995), Brabyn (1997), Giles (1998), and Burrough et al. (2001), have used computed topographic attributes to generate formal landform classifications.

## 4. Secondary Topographic Attributes

The secondary attributes that are computed from two or more primary attributes are important because they offer an opportunity to describe pattern as a function of process (Table 2). Those attributes that quantify the role played by topography in redistributing water in the landscape and in modifying the amount of solar radiation received at the surface have important hydrologic, geomorphic, and ecological consequences in many landscapes. These attributes may affect soil characteristics (because the pedogenesis of the soil catena is affected by the way water moves through the environment in many landscapes), distribution and abundance of soil water, susceptibility of landscapes to erosion by water, and the distribution and abundance of flora and fauna (Wilson and Gallant 2000). The methods used to calculate three sets of compound topographic indices that have found widespread use in a variety of hydrologic, geomorphic, and ecological applications are described below.

Two topographic wetness indices have been used extensively to describe the effects of topography on the location and size of saturated source areas of runoff generation as follows:

$$W_T = \ln (A_s / T \tan \beta) \quad (1)$$

$$W = \ln (A_s / \tan \beta) \quad (2)$$

where $A_s$ is the specific catchment area ($m^2 m^{-1}$), $T$ is the soil transmissivity when the soil profile is saturated, and $\beta$ is the slope gradient (in degrees) (Moore et al. 1991, 1993c). The second equation contains one less term because it assumes uniform soil properties (i.e. that the soil transmissivity is constant throughout the landscape). Wood et al. (1990) have shown that the variation in the topographic component is often far greater than the local variability in soil transmissivity and that Equation 2 can be used in place of Equation 1 in many landscapes. Both of these indices predict that points lower in the catchment, and particularly those points near the outlets of the main channels, are the wettest points in the catchment, and the soil water content decreases as the flow lines are retraced upslope to the catchment divide (Figure 2) (Wilson and Gallant 1998).

**Table 2.** Secondary topographic attributes that can be calculated from DEM data (from Wilson and Gallant 2000)

| **Attribute** | **Definition** | **Significance** |
|---|---|---|
| Topographic wetness indices | $W_T = \ln\left(\frac{A_s}{T \tan \beta}\right)$ | This equation assumes steady state conditions and describes the spatial distribution and extent of zones of saturation (i.e. variable source areas) for runoff generation as a function of upslope contributing area, soil transmissivity, and slope gradient. |
| | $W = \ln\left(\frac{A_s}{\tan \beta}\right)$ | This particular equation assumes steady state conditions and uniform soil properties (i.e. transmissivity is constant throughout the catchment and equal to unity). This pair of equations predicts zones of saturation where $A_s$ is large (typically in converging segments of landscapes), β is small (at base of concave slopes where slope gradient is reduced), and $T_i$ is small (on shallow soils). These conditions are usually encountered along drainage ways and in zones of water concentration in landscapes. |
| | $W = \ln\left(\frac{A_s}{\tan \beta}\right)$ | This quasi-dynamic index substitutes effective drainage area for upslope contributing area and thereby overcomes limitations of steady-state assumption used in first pair of equations. |
| Stream power indices | $SPI = A_s \tan \beta$ | Measure of erosive power of flowing water based on assumption that discharge (*q*) is proportional to upslope contributing area ($A_s$). Predicts net erosion in areas of profile convexity and tangential concavity (flow acceleration and convergence zones) and net deposition in areas of profile concavity (zones of decreasing flow velocity). |
| | $LS = (m+1)\left(\frac{A_s}{22.13}\right)^m \left(\frac{\sin \beta}{0.0896}\right)^n$ | The sediment transport capacity index was derived from unit stream power theory and is equivalent to the length-slope factor in the Revised Universal Soil Loss Equation in certain circumstances. Another form of this equation is sometimes used to predict locations of net erosion and net deposition areas. |
| | $CIT = A_s(\tan \beta)^2$ | Variation of stream power index sometimes used to predict the locations of headwaters of first-order streams (i.e. channel initiation). |

| | | |
|---|---|---|
| Radiation indices | $R_t = (R_{th} - R_{dh})F + R_{dh}v + R_{th}(1-v)\alpha$ | This equation estimates the total shortwave irradiance incident at the earth's surface for some user-defined period ranging in length from one day to one year. The three main terms account for direct-beam, diffuse, and reflected irradiance. A variety of methods are used by different authors to calculate these individual components. The methods vary tremendously in terms of sophistication, input data, and accuracy. |
| | $L_{in} = \varepsilon_a \sigma T_a^4 v + (1-v)L_{out}$ | This equation estimates the incoming or atmospheric long-wave irradiance. |
| | $L_{out} = \varepsilon_s \sigma T_s^4$ | This equation estimates the outgoing long-wave irradiance. |
| | $R_n = (1-\alpha)R_t + \varepsilon_s L_{in} - L_{out}$ | This equation estimates the net radiation or surface energy budget at the earth's surface for some user-defined period. May or may not account for the effects of clouds depending on the methods and data sources used to estimate individual short-wave radiation components. |
| Temperature indices | $T = T_b - \frac{T_{lapse}(Z - Z_b)}{1000} + CS\left(1 - \frac{LAI}{LAI_{\max}}\right)$ | This equation is used to extrapolate minimum air, maximum air, and surface temperatures for a nearby climate station to other parts of the landscape. This equation corrects for elevation via a lapse rate, slope-aspect effects via the short-wave radiation ratio, and vegetation effects via a leaf area index. |

These indices are used in TOPMODEL (Beven and Kirkby 1979) to characterize the spatial distribution and extent of zones of saturation and variable source areas for runoff generation. O'Loughlin (1986) also used these indices to identify surface saturation zones in landscapes. Burt and Butcher (1986), Jones (1986), and Moore et al. (1988) used variants of these compound topographic wetness indices to describe the spatial distribution of soil water content. Moore et al (1986) showed how the wetness index versus percent saturated source area relationship can be combined with observed streamflow data and used to estimate the effective transmissivity of a small forested catchment. Sivapalan et al. (1987) used this index to characterize hydrologic similarity and Phillips (1990) used it to delineate wetlands in a coastal plain drainage basin. Moore et al. (1993a, b) used slope and topographic wetness index to characterize the spatial variability of soil properties for a toposequence in Colorado. Montgomery and Dietrich (1995) used TOPOG (O'Loughlin 1986) and this relative measure of saturation to analyze the stability of each topographic element for the case of cohesionless soils of spatially constant thickness and saturated conductivity in three California, Oregon, and Washington study areas.

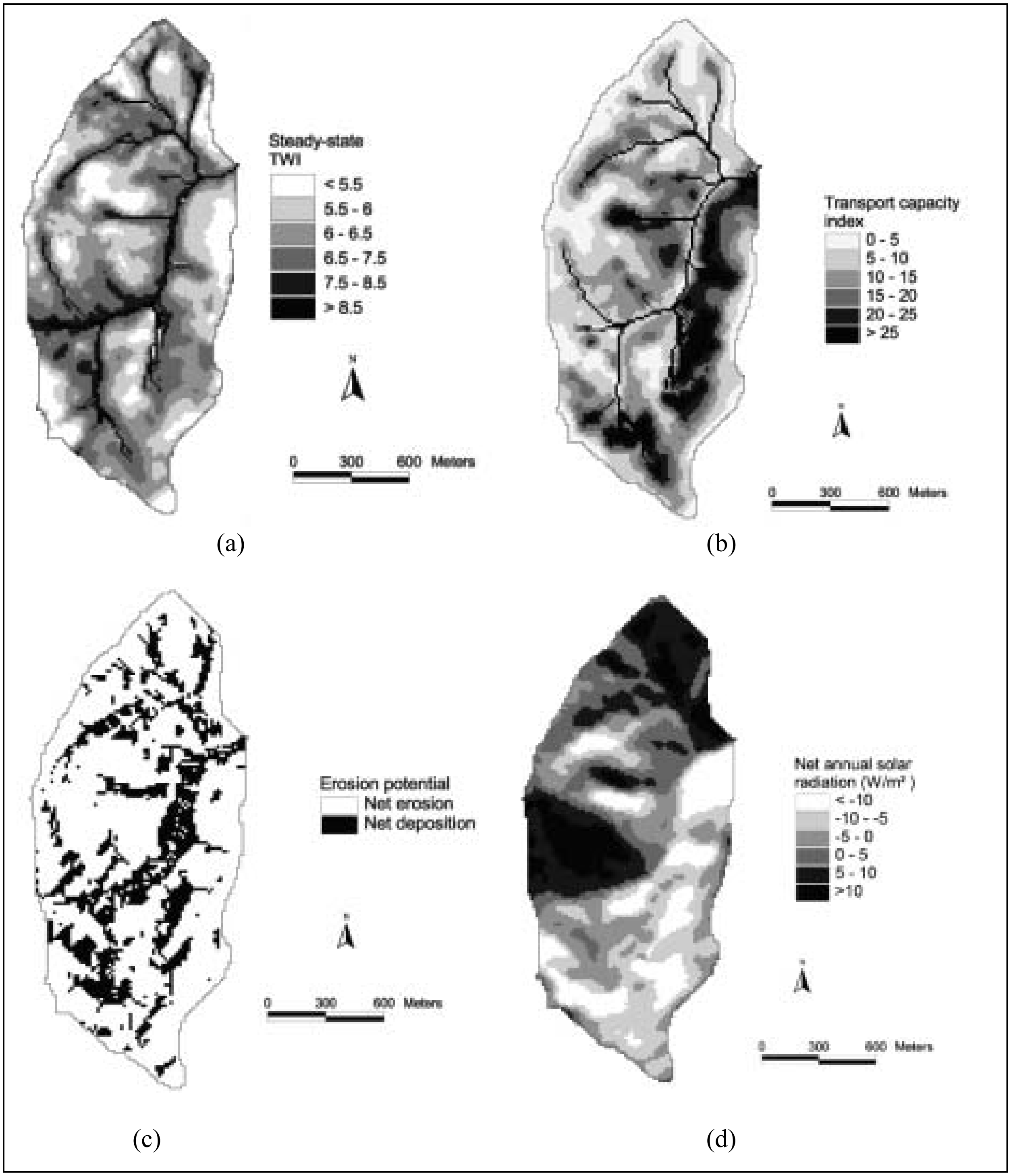


**Figure 2.** Maps of Cottonwood Creek, Montana showing (a) steady-state topographic wetness index, (b) sediment transport capacity index, (c) change in sediment transport capacity index, and (d) net annual solar radiation

These types of static indices must be used carefully to predict the distribution of dynamic phenomena like soil water content because surface saturation is a threshold process and because of hysteretic effects (Burt and Butcher 1986, Moore et al. 1991). In addition, there are several important and implicit assumptions in the derivation of the two wetness indices described above. Most notably, the gradient of the piezometric head, which dictates the direction of subsurface flow, is assumed to be parallel to the land surface and steady-state conditions are assumed to apply (Moore et al. 1993c). Several authors have described the pitfalls of using these indices in inappropriate ways. Jones (1986, 1987), for example, discussed the advantages and limitations of wetness indices as indicators of spatial patterns of soil water content and drainage. Quinn et al. (1995) summarized various problems and described how steady-state topographic wetness indices can be calculated and used effectively in TOPMODEL. Barling et al. (1994) and Wood et al. (1997) have proposed alternative forms of these equations in hopes of overcoming some of these shortcomings.

Several terrain-based stream power and sediment transport capacity indices have also been proposed (Table 2). Stream power is the time rate of energy expenditure and has been used extensively in studies of erosion, sediment transport, and geomorphology as a measure of the erosive power of flowing water (Moore et al. 1991). It is usually computed as:

$$\Omega = \rho g q \tan\beta \qquad (3)$$

where $\rho g$ is the unit weight of water, $q$ is the discharge per unit width, and $\beta$ is the slope gradient (in degrees). The compound topographic index $A_s \tan\beta$ is, therefore, a measure of stream power since $\rho g$ is essentially constant and $q$ is often assumed to be proportional to $A_s$. Several researchers have used variations of this index to predict the locations of ephemeral gullies. Thorne et al. (1986) multiplied this index by plan curvature and predicted both the locations of ephemeral gullies and the cross-sectional areas of the gullies after one year of development with variants of this new index. Moore et al. (1988) showed that ephemeral gullies formed where $W > 6.8$ and $A_s \tan\beta > 18$ for a small semi-arid catchment in Australia, and Srivastava and Moore (1989) found that ephemeral gullies formed where $W > 8.3$ and $A_s \tan\beta > 18$ on a small catchment in Antigua. Moore et al. (1991) concluded that threshold values of these indices are likely to vary from place to place because of differences in soil properties. Moore and Nieber (1989) used the stream power index to identify places where soil conservation measures that reduce the erosive effects of concentrated flow, such as grassed waterways, should be installed. Montgomery and Dietrich (1989, 1992) and Montgomery and Foufoula-Georgiov (1993) used a variation of this index ($A_s (\tan\beta)^2$) to predict the headwaters of first order streams (i.e. the locations of channel initiation).

A second compound index was derived by Moore and Burch (1986a, b, c) from unit stream power theory and a variant used in place of the length-slope factor in the Revised Universal Soil Loss Equation (RUSLE) for slope lengths < 100 m and slopes < 14$^{o}$ as follows:

$$LS = (m + 1)\,(A_s / 22.13)^m\,(\sin\beta / 0.0896)^n \qquad (4)$$

where $m = 0.4$ and $n = 1.3$ (Moore and Wilson 1992, 1994). Both this and the next equation are nonlinear functions of slope and specific discharge (Figure 2). This new index calculates a spatially distributed sediment transport capacity and may be better suited to landscape assessments than the original empirical equation because it explicitly accounts for flow convergence and divergence (Moore and Wilson 1992, Desmet and Govers 1996b).

A third sediment transport capacity index has been proposed to differentiate net erosion and net deposition areas:

$$\Delta T_{cj} = [A_{sj-}^{\ m} (\sin \beta_{j-})^n - A_{sj}^{\ m} (\sin \beta_j)^n] \qquad (5)$$

where $\varphi$ is a constant, subscript $j$ signifies the outlet of cell $j$, and subscript $j$- signifies the inlet to cell $j$ (Moore and Wilson 1992, 1994). This index will predict erosion in areas experiencing an increase in sediment transport capacity and deposition in areas experiencing a decrease in sediment transport capacity (see Figure 2 for an example).

Mitasova et al (1996) implemented variants of these equations in the GRASS GIS (U.S. Army Corps Engineers 1987) and showed how net erosion areas coincided with areas of profile convexity and tangential concavity (flow acceleration and convergence) and net deposition areas coincided with areas of profile concavity (decreasing flow velocity). These patterns match those observed by Martz and De Jong (1987), Foster (1990), Sutherland (1991), and Busacca et al (1993) in a variety of landscapes.

Several authors have criticized the use of Equation 4 in place of the original slope gradient and length terms in RUSLE (Renard et al. 1991) and its predecessors. Interested readers should consult Moore and Wilson (1992, 1994), Foster (1994), Mitasova et al. (1996, 1997), and Desmet and Govers (1996b, 1997) for additional details. Wilson and Lorang (1999) recently summarized the key elements of this debate, and why the terrain-based approach of Mitas et al. (1996) probably represents a superior approach for simulating the impact of complex terrain and various soil and land cover changes on the spatial distribution of soil erosion and deposition.

The third and final set of compound indices is used to estimate the spatial and temporal distribution of solar radiation at the earth's surface. Topography may exert a large impact on the amount of solar energy incident at a location on the earth's surface (Dubayah and Rich 1995). Variations in elevation, slope, aspect, and local topographic horizon can cause substantial differences in solar radiation and thereby affect air and soil heating, evapotranspiration, and primary production (Gates 1980, Linacre 1992). These processes may, in turn, affect the distribution and abundance of flora and fauna (see Hutchins et al. 1976; Kirkpatrick and Numez 1980; Austin et al. 1983, 1984; Tajchman and Lacey 1986; Noguchi 1992a, b; and Moore et al. 1993d for examples of studies showing that the distributions of solar radiation and vegetation are highly correlated).

Numerous approaches have been proposed to calculate radiation fluxes and temperature indices. Most of the radiation models incorporate one or more of the following equations. The net radiation, $R_n$, received by an inclined surface can be written as:

$$R_n = (1 - \alpha)(R_{dir} + R_{dif} + R_{ref}) + \varepsilon_s L_{in} - L_{out} = (1 - \alpha) R_t + L_n \quad (6)$$

where $\alpha$ is the surface albedo, $\varepsilon_s$ is the surface emissivity, $R_{dir}$, $R_{dif}$, and $R_{ref}$ are the direct, diffuse, and reflected short wave irradiance, respectively, for which $R_t = R_{dir} + R_{dif} + R_{ref}$, the global short wave irradiance, $L_{in}$ is the incoming or atmospheric long wave irradiance, $L_{out}$ is the outgoing or surface long wave irradiance, for which $\varepsilon_s L_{in} - L_{out} = L_n$, the net long wave irradiance.

The total short wave irradiance is estimated by:

$$R_t = (R_{th} - R_{dh}) F + R_{dh} \upsilon + R_{th} (1 - \upsilon) \alpha \quad (7)$$

where $R_{th}$ and $R_{dh}$ are the total and diffuse radiation on a horizontal surface, and $F$ is the potential solar radiation ratio, which is the ratio of the potential solar radiation ($R_o$) on a sloping surface to that on a horizontal surface ($R_{oh}$), and $\upsilon$ is the skyview factor, which is the fraction of the sky that can be seen by the sloping surface. The total and diffuse short wave irradiances on a horizontal surface are often expressed as functions of the total and diffuse transmittances of the atmosphere and the potential solar radiation on a horizontal surface. These transmittances are functions of the thickness and composition of the atmosphere, such as the water vapor, dust, and aerosol content (Lee 1978, Gates 1980).

The long wave irradiance components can be approximated on a cell-by-cell basis using:

$$L_{out} = \varepsilon_s \sigma T_s^4 \quad (8)$$
$$L_{in} = \varepsilon_a \sigma T_a^4 \upsilon + (1 - \upsilon) L_{out} \quad (9)$$

where $\varepsilon_a$ is the atmospheric emissivity (a function of air temperature, vapor pressure and cloudiness), $\sigma$ is the Stefan-Boltzman constant, $T_s$ is the mean surface temperature, and $T_a$ is the mean air temperature. Two complementary equations that utilize modifications of a simple approach proposed by Running et al. (1987), Hungerford et al. (1989), and Running (1991) for estimating the spatial distribution of minimum, maximum, and average air temperature are summarized in Table 2.

The different indices that have been proposed vary in terms of the methods, data sources, and assumptions used to estimate individual components. Hence, Wilson and Gallant (2000) have described an approximate method for estimating each of the above fluxes at any location in a topographically heterogeneous landscape. The variation in potential solar radiation can be estimated over a catchment as a function of slope, aspect, topographic shading and time of year, and then adjusted for cloud, atmospheric, and land cover effects with this method. The variables that serve as model inputs, such as albedo, cloudiness, emissivity, sunshine fraction, mean air and surface temperatures and clear sky transmittances, can be varied on a monthly or annual basis (Wilson and Gallant 2000). A sample map produced with this method for a small catchment in southwest Montana is shown in Figure 2.

Hetrick et al. (1993a, b) used latitude, atmospheric transmissivity, slope, aspect, topographic shading and time of year to estimate direct and diffuse irradiance at each grid point. The effects of cloud cover, which are likely to be substantial in many humid environments, were not accounted for in this model. Kumar et al. (1997) chose a simpler approach and used latitude and a series of topographic attributes derived from a square-grid DEM to estimate clear sky direct beam short wave radiation. These relationships are generally straightforward and numerous authors have summarized the appropriate equations for both horizontal and sloping sites (see Lee 1978, Gates 1980, Iqbal 1983, and Linacre 1992 for additional details). Most of the challenges (problems) are encountered when atmospheric effects (precipitable water, dust, etc.), cloud cover, and land surface characteristics (albedo) are considered. Dubayah and Rich (1995) have reviewed many of the important computational challenges and errors that are likely to be encountered in building accurate, physically-based topographic solar radiation models.

## 5. Conclusions

Although it is not always apparent to users of terrain analysis, the three sets of indices described above are simplified process models and are not applicable in all situations (Wilson and Gallant 1998). The topographic wetness index, for example, is based on the assumption that the soil hydraulic conductivity decreases exponentially with depth so that subsurface flow is confined to a shallow layer. If this is not the case, topographic wetness indices will be poor predictors of the spatial distribution of soil water. An alternative index might be developed to better represent the topographic effect on water distribution, perhaps based on groundwater potential expressed as a simple elevation difference above a local mean or minimum (e.g. Hinton et al. 1993).

The topographic indices introduced on the preceding pages account for the component of the spatial variability of processes that is due to topographic effects. Other spatially variable factors are usually involved, such as soil hydraulic properties and vegetation in the case of soil water. In some instances, the spatial variations in these other attributes are themselves linked to the topographic indices. The spatial variability of soil properties is one case where significant links have been established (e.g. Moore et al. 1993e, Wilson et al. 1994). There are other properties though where explicit incorporation of the spatial variation of other important components of process models would substantially improve the predictive accuracy of topographic indices particularly when working at a broad landscape scale as opposed to the small catchment scale. Surficial geology and, in some cases, climate are likely candidates for inclusion in these types of applications.

Additional problems may be encountered by the terrain analyst or user because the spatial and statistical distributions of the computed primary and secondary topographic attributes may be affected by the presence of errors in the source data and the choice of computer algorithm and/or element size (i.e. grid spacing). The most serious problems are usually encountered when secondary attributes are derived: the topographic wetness and sediment transport capacity indices are very sensitive to the presence of errors in source (elevation) data in flat areas and to the choice of flow routing algorithm (Moore et al 1993e). The identification of problems with existing algorithms and fundamental role of flowing water in controlling or explaining many key environmental processes and patterns are likely to promote further methodological innovation in this area.

Taken as a whole, these continuously varying but gridded terrain attributes demonstrate how simple spatial models can be combined with qualitative reasoning to improve our understanding and management of environmental systems (Wilson and Burrough 1999). In addition, most of the current methods and data sources work best at intermediate spatial (hillslopes and catchments) and temporal scales (measured in terms of months or years). These approaches and scales are consistent with our current ability to represent biophysical systems and the availability of data and as such, they are likely to increase the quantity and quality of the scientific concepts and geospatial information used in environmental assessments and management applications in the immediate future.

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