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ABSTRACT
Stream power–based models of bedrock landscape development are effective at producing synthetic topography with realistic fluvial-network topology and three-dimensional topography, but they are difficult to calibrate. This paper examines ways in which field observations, geochronology, and digital elevation model (DEM) data can be used to calibrate a bedrock landscape development model for a specific study site. We first show how uplift rate, bedrock erodibility, and landslide threshold slope are related to steady-state relief, hypsometry, and drainage density for a wide range of synthetic topographies produced by a stream power–based planform landscape development model. Our results indicate that low uplift rates and high erodibility result in low-relief, high drainage density, fluvially dominated topography, and high uplift rates and low erodibility leads to high-relief, low drainage density, mass wasting–dominated topography. Topography made up of a combination of fluvial channels and threshold slopes occurs for only a relatively narrow range of model parameters. Using measured values for hypsometric integral, drainage density, and relief, quantitative values of bedrock erodibility can be further constrained, particularly if uplift rates are independently known.

We applied these techniques to three sedimentary rock units in the western Transverse Ranges in California that have experienced similar climate, uplift, and incision histories. The 10Be surface exposure dating and optically stimulated luminescence (OSL) burial dating data indicated that incision of initially low-relief topography there occurred during the last ~60 k.y. We estimated the relative dependence of drainage area and channel slope on erosion rate in the stream power law from slope-area data, and inferred values for bedrock erodibility ranging from 0.09 to 0.3 m^{0.2-0.6/k.y.} for the rock types in this study area.

Keywords: landscape development, erosion, modeling, DEM, Cuyama Basin.

INTRODUCTION
Numerical landscape development models are useful tools for understanding how past surficial processes have led to current landscape form. By simulating surficial processes in a landscape development model and creating synthetic landscape forms that mirror reality, we may develop quantitative understanding of landscape-altering processes. By understanding the roles of each parameter in such a model, we can better understand the importance of each modeled process. This can also increase our ability to make specific predictions about how future surface process dynamics may lead to changes in landscape form.

Stream power or shear stress fluvial bedrock erosion laws form the foundation for many bedrock landscape development models (e.g., Howard, 1994; Whipple and Tucker, 1999). When fluvial bedrock channel development models are coupled with hillslope process models that include threshold landsliding, slope diffusion, or other reasonable hillslope process components, three-dimensional landscape development modeling is possible (e.g., Tucker et al., 2001; Howard, 1994). We were motivated by the need to understand how each parameter in bedrock landscape development models affects model output topography, and the need to develop general techniques for calibrating landscape development models using geologic and morphometric analyses. This motivation led us to attempt to calibrate a landscape development model with a minimum of free parameters using geologic and morphologic data from a field site in southern California.

One of the simplest mathematical foundations for bedrock landscape development modeling is the stream power law for fluvial erosion:

\[
\frac{dh}{dt} = U - KA^\alpha \left( \frac{dh}{dx} \right)^n,
\]

in which \( h \) is elevation; \( t \) is time; \( U \) is rock uplift rate and/or local base-level lowering rate; \( K \) is a constant often referred to as the “coefficient of bedrock erodibility,” which is inversely proportional to the landscape’s resistance to erosion (larger values of \( K \) are generally related to weaker rocks); \( A \) is the contributing drainage area that serves as proxy for local discharge, which, as it increases, leads to higher fluvial erosion rates; \( x \) is the along-channel distance used along with \( h \) to calculate local slope, which, as it increases, leads to increased fluvial energy and therefore erosion rates; and \( m \) and \( n \) are exponents (Whipple and Tucker, 1999) that control the relative dependence of channel slope and discharge on local erosion rate.

Stock and Montgomery (1999) proposed a range in \( K \) over five orders of magnitude for varying rock types, and this wide range has been used in other modeling studies. Because the stream power law is very sensitive to the value of \( K \) and because Snyder et al. (2000) proposed a linkage between uplift rate and \( K \), we were interested in creating a more specific calibration technique for the stream power law that relies on...
geologic constraints of uplift rate and morphometric landscape analyses to calibrate $K$. Snyder et al. (2000) also provided insight regarding use of landscape morphometry to constrain the values of stream power law exponents $m$ and $n$. By integrating these studies’ findings into a fully coupled landscape modeling environment, we hoped to further refine our understanding of the effect of model parameters as a step toward improving our ability to calibrate even more sophisticated landscape development models.

We utilized the model of Pelletier (2004) and field and geochronological data from the upper Cuyama Valley in southern California in an effort to develop a general method for calibrating the coefficient of bedrock erodibility $K$ in the stream power law. Our model erodes bedrock channels according to Equation 1 and incorporates mass wasting–dominated hillslope development by removing hillslope material during each time step from cells that exceed a local landslide-threshold slope, $S_t$. We suggest that because the model is limited to one of the simplest possible formulations of a coupled hillslope-channel landscape development model that still appears to produce physically reasonable landscapes (i.e., landscapes made up of slopes and channels in which channel network topology, channel longitudinal profiles, and the general appearance of the landscape appears similar to actual landscapes), we are able to uniquely determine each parameter’s effect on landscape morphology. We do not suggest that the model adequately captures the possible range of geomorphic processes at work on uplifting orogenic blocks.

In the first part of this study, we performed model runs in which we varied the three key model parameters $K$, $U$, and landslide-threshold slope, $S_t$, systematically in order to evaluate their influence on synthetic landscape morphometry. For each set of parameters, we ran our model until steady state was achieved and then extracted values for (1) landslide threshold slope; (2) hypsometric integral (i.e., the area beneath the curve that relates the percentage of area to cumulative percentage of area) (Strahler, 1952); (3) drainage density (i.e., the proportion of the landscape that is occupied by a fluvial channel); (4) topographic relief; (5) mean elevation; and (6) the nondimensional “ruggedness number,” which is equal to topographic relief multiplied by drainage density (Melton, 1957). By carefully controlling model parameters and characterizing resulting topographies, we were able to quantify how each model parameter affects output topology.

Next, we calibrated the parameter $K$ for three distinct lithologic units in southern California by comparing model-produced synthetic morphometry against actual morphometry measured in the field and from digital elevation model (DEM) data. Geologic and geochronological data provide important constraints on this work. Incision into an alluvium-capped late Pleistocene erosion surface has formed equal-aged drainage networks that have varying morphology in several distinct rock types in the upper Cuyama Valley, California (Davis, 1983). This study area is therefore appropriate for this type of study because initial topography, age of incision, downcutting rate, topographic relief, and landslide threshold slope are all known and vary across three lithologies. We characterized real-world topography and compared it to our model-run outputs generated with controlled input parameters. From this, we estimated best-fit values of $K$ using data from digital elevation model analysis and geochronology. Our results apply to drainage basins that have fluvially dominated channel incision and planar landside-dominated hillslopes.

**Previous Bedrock Landscape Development Model Calibration**

Howard (1997) used a shear stress–based landscape development model (introduced in Howard, 1994) in an effort to detail how model parameterizations affected landscape morphology results and to interpret the morphology and development of badlands in Utah. This work was an important step in comparison of numerical landscape development models to actual topography and highlighted the need for additional studies of its kind. Howard considered threshold landsliding and diffusive hillslope development for both threshold and nonthreshold shear stress–based fluvial incision. He then applied a best-fit model choice, based primarily on relief production, to support interpretations about the erosional history of badlands near Caineville, Utah. This calibration was semiquantitative, and model run topography did not duplicate actual drainage density owing to computational limitations at the time. He did not undertake a comprehensive set of steady-state model runs to determine each model parameter’s effect on model topography, but rather focused on identifying the most likely temporal incision regime. He used a one-dimensional formulation to assess model characterization effect on drainage density. This study serves as an excellent example of the usefulness of field-calibrated landscape development modeling, and it serves as an important complement to the work presented here.

Whipple and Tucker (1999) provided an excellent review and analysis of the fundamentals of stream power law modeling. Attempts to calibrate the stream power law have relied primarily on analysis of “paleo-” and modern stream profiles in a two-dimensional sense. These studies demonstrated the applicability of stream power law–based techniques to understanding landscape development. Howard and Kerby (1983) first formalized ideas about the relationships between discharge, slope, and bedrock resistance into a stream power law and estimated values for $m$ and $n$ using slope-area relationships in a badland channel. Following their general techniques, Stock and Montgomery (1999), Seidl et al. (1994), and Rosenberg and Anderson (1994) each calculated values for $K$ using data from channel longitudinal profiles. In a particularly influential contribution, Stock and Montgomery (1999) integrated topographic data from varied settings in a consistent manner to provide several order-of-magnitude estimates of $K$ for various rock types using a number of stream power law formulations. Their study suggested that values for $K$ vary over as much as five orders of magnitude in different lithologies.

Snyder et al. (2000) highlighted the possible dependence of $K$ on local values for uplift rate $U$ by analyzing stream profiles in California that had varying uplift rates as constrained by uplifting marine platforms. They found that, depending on the choice of exponent $n$, values for $K$ could increase by as much as sixfold in areas of higher uplift rate, even in areas with very similar lithology. However, the processes responsible for a relationship between $U$ and $K$ have not been determined.

Use of the stream power law to model bedrock incision does not take into account all details of fluvial process, and more sophisticated models have been produced to improve on the stream power law. These refinements have been made within several landscape development models, usually in an effort to illuminate details of specific (though often common) processes at work in specific field sites. These refinements have included debris-flow incision of channels (Stock and Dietrich, 2003), saltation and abrasion-forced bedrock channel incision (Sklar and Dietrich, 2004), soil production, slope cohesion, and non-mass-movement–dominated hillslopes (e.g., Roering, et al. 2001). Perhaps the most well-developed model is the Channel-Hillslope Integrated Landscape Development (CHILD) model (e.g., Tucker et. al., 2001), which is a flexible and sophisticated landscape development model that includes many parameters to help simulate many specific surface processes. Models such as these will undoubtedly form the foundation of the next generation of three-dimensional landscape development models. More sophisticated models have a larger number of parameters in need of calibration, but as we become more confident in our ability to calibrate specific mechanistic process-models, we
can apply the techniques proposed in this study to better understand each parameter’s effect on large-scale topographic indicators. We believe that while our study does not calibrate the most sophisticated landscape development models available, it is an important step in our ability to do so.

MODELING APPROACH

Most landscape development models to date have been designed such that an ideal steady state is reached when exhumation equals uplift; this leads to a time-invariant topographic configuration. Recently, Pelletier (2004) showed that important long-term variation in landscape development can be captured with incorporation of a more sophisticated flow-routing algorithm. Pelletier’s model followed upon classic models developed over the past two decades (e.g., Willgoose et al., 1991; Howard, 1994; Tucker and Slingerland, 1997). For the purpose of determining whether erosional parameters can be determined from topography with a forward-modeling approach, Pelletier’s model has the advantage of requiring very few input parameters.

Following Pelletier (2004), we utilized a bedrock landform development model that solves Equation 1 on a rectangular grid that is subject to a uniform rate of combined tectonic uplift and base-level lowering and a fixed-elevation boundary condition (all material that reaches the edge of the grid is removed in each time step). The model uses bifurcation-flow routing as a technique for calculating parameter \( A \). This allows for long-term migration of ridges and valleys, and better approximates discharge-dependent incision rates in low-relief areas such as hillslopes and valley floors by allowing for divergent flow and thalweg widths wider than model pixel size. The model includes threshold landsliding such that material from slopes steeper than the defined threshold slope, \( S_c \), is removed from upslope each time step until all slopes do not exceed \( S_c \). A schematic representation of model behavior is presented in Figure 1.

The parameters needed for each model run are grid geometry, \( U \), \( K \), \( S_c \), \( m \), and \( n \), and model run time. For calibration runs, \( U \) and \( S_c \) were approximated from geologic and topographic data, and \( K \) was our primary independent variable. While the precise values of \( m \) and \( n \) are difficult to constrain, determining \( mn \) is relatively straightforward by regression of slope and area data (Tucker and Whipple, 1999; Snyder et al., 2000). For the model runs performed in the effort to assess the effects of variations in \( K \), \( U \), and \( S_c \) on model topography, we assumed \( n = 1 \) and \( m = 0.5 \), which are commonly used values for these parameters. For calibration efforts, we determined \( mn \) from slope and area data, and assumed \( n = 1 \) (following Snyder et al., 2000).

Characterization of Synthetic and Real Topography

In order to compare model results with geological reality, we characterized both model-derived synthetic topography and upper Cuyama River drainage basin topography according to several morphometric indicators. For characterization of real topography, we used 10 m U.S. Geological Survey (USGS) DEM data and direct field observation. We used the following parameters to characterize both model topography and real topography: (1) landslide threshold slope, which was measured in the field with an inclinometer (slopes in the field area are dominantly not soil mantled or thin soil mantled, planar, and of fairly consistent magnitude within a given lithology, and they were measured in several locations per lithology with a Brunton inclinometer, and validated using contour map and DEM analyses); (2) hypsometric integral; (3) drainage density; (4) topographic relief; (5) mean elevation; and (6) the nondimensional “ruggedness number,” which equals relief multiplied by drainage density. Modeled drainage density was calculated by characterizing all parts of the grid that had or exceeded landslide threshold slopes as being hillslope, and all others as being part of the channel network. In our synthetic topography, landscapes were made up of only slopes and channels; so all nonslope pixels were, by definition, channels. The drainage density was then taken as the total length of channels divided by total basin area (with units of I/L). We normalized this to the maximum possible drainage density in order to minimize dependence on DEM pixel size. This method works well for idealized model outputs, but in actual topography variation in threshold slope, the presence of colluvial material and other topographic irregularities leads to parts of the landscape being inappropriately lumped in with the channel network, which leads to an overestimation of drainage density. For this reason, we estimated the maximum channel slope within each lithologic unit from field and aerial photo observations of channel head positions and map contour measurements. We applied that measurement as the upper slope threshold for classification of cells as channels and all steeper areas as hillslopes. While this method also is not perfect, we believe it to be a reasonable approach for comparing synthetic and real topography.

RESULTS: SYNTHETIC LANDSCAPE DEVELOPMENT

To capture the full range of possible model and real-world topographies, we used a wide range of model parameters in our synthetic model runs, using 15 km² grids and 30 m grid cells. In order to evaluate the calculated variability in \( K \) reported by Stock and Montgomery (1999), we varied \( K \) over five orders of magnitude ranging from \( 10^{-4} \) to \( 10^4 \) k.y.\(^{-1}\). We varied landslide-threshold slope \( S_c \) from 20° to 40° and \( U \) from 0.1 m/k.y. to 50 m/k.y. For these runs,
Model Behavior and Classification of Model Topography

Model run results were synthetic landscapes that were classified into three categories. These categories are best applied in model-space, because actual topography has more complexity than this classification scheme can adequately capture. Type I synthetic topographies have dendritic drainage network morphology in which fluvial erosion dominates the entire landscape such that all slopes are below the threshold angle $\theta$ and topographic relief is limited. Type II synthetic topographies consist of a combination of dendritic channel networks and planar threshold slopes. Type III synthetic topographies are characterized by threshold landsliding that outpaces fluvial erosion such that the steady-state form is made up of entirely unchanneled threshold slopes. Figure 2 illustrates the three types of synthetic topography that result from varying $K$ while holding other model parameters constant. Table 1 contains the key morphologic indicators extracted from those model runs. Type III topographies are clearly not present in their idealized pyramidal form in nature but perhaps roughly approximate mass wasting-dominated bedrock plateaus. Type III topographies are likely limited in nature by the tendency for large topographic loads to cause subsidence, which would offset higher values of $U$, especially in areas of resistant bedrock. Furthermore, resistant, rapidly uplifting masses are clearly not well modeled by our simple model because the model does not incorporate other processes such as glacial erosion and channel debris flows, which are clearly important in many high-elevation orogenies worldwide. Type II topographies are common and simulate the geologic reality in a number of tectonically active regimes, including the upper Cuyama drainage basin. Type I topographies generally form low relief (usually much less than 100 m) and contain drainage divides that are eroded primarily by fluvial processes. These are perhaps similar to fluvially dominated badland topography, but are likely highly transient over geological time scales in natural settings. Figure 3 shows that for all model runs ($n = 172$), type II basins only form for a relatively narrow range of values for $K$ and $U$. For a given uplift rate $U$, type II topography occurs over less than three orders of magnitude of $K$.

The limits of applicability of this model, and our wide-ranging model-run output data set, deserve clear discussion. Since our model is limited in the number of processes explicitly represented, the end-member model topographies are the least appropriate for such a simple model. Low-relief (generally less than a few hundred meters of relief), low rock uplift rate (generally less than one meter per thousand years) topographies are less likely to undergo threshold landsliding. In these less active landscapes, processes related to soil-formation, creep, and diffusive processes are likely important. Additionally, in areas of high uplift rate, glacial and/or debris-flow processes likely become more important. To put our range of model uplift rates in perspective, some of the highest uplift rates known currently are on the order of >10 m/k.y. from places such as the Himalayas (e.g., Kirby and Whipple, 2001).

Sobel et al. (2003) also showed that in rapidly uplifting zones that have resistant lithologies, fluvial-system development can become “defeated,” leading to formation of internal drainage. This phenomenon is not represented in our model and requires the explicit representation of distinct structural boundaries. This model is most applicable to moderate-relief, moderate uplift rate topography in which (as in our field area) field inspection indicates that threshold landsliding and fluvial channel erosion dominate. We present the full data set as a starting point for analyses of this type, and caution the reader against applying it universally without accounting for process variation.

In order to determine how morphologic indicators depend on variables $U$, $K$, and $S_e$, we characterized our synthetic topography according to the morphologic indicators mean elevation, topographic relief, drainage density, ruggedness number, and hypsometric integral. In our method of calculating drainage density, type I topographies on 30 m grids achieved a maximum theoretical drainage density of 0.0333 m$^{-1}$ (sum of model channel pixel

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$K = 0.0001$ k.y.$^{-1}$

$K = 0.005$ k.y.$^{-1}$

$K = 0.001$ k.y.$^{-1}$

$K = 0.01$ k.y.$^{-1}$

$K = 0.1$ k.y.$^{-1}$

Figure 2. Oblique shaded-relief images of synthetic topography formed by varying only model parameter $K$. All model runs are $15 \times 15$ km, with 30 m pixel size. Uplift rate is 1 m/k.y., threshold slope is 0.67, $m = 0.5$, and $n = 1$. See Table 1 for output morphologic indicators.

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$K = 0.0001$ k.y.$^{-1}$

$K = 0.005$ k.y.$^{-1}$

$K = 0.001$ k.y.$^{-1}$

$K = 0.01$ k.y.$^{-1}$

$K = 0.1$ k.y.$^{-1}$

GSA Data Repository item 2007003, landscape development model data, is available on the Web at http://www.geosociety.org/pubs/ft2007.htm. Requests may also be sent to editing@geosociety.org.
TABLE 1. MORPHOLOGIC INDICATORS FROM TOPOGRAPHY WITH VARYING K VALUES

<table>
<thead>
<tr>
<th>K (k.y.⁻¹)</th>
<th>U (m/k.y.)</th>
<th>S_c ¹</th>
<th>Mean elevation (m)</th>
<th>Relief (m)</th>
<th>Normalized drainage density</th>
<th>Ruggedness</th>
<th>Hypsometric integral</th>
<th>Topography type</th>
<th>U/K</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.0001</td>
<td>1</td>
<td>0.67</td>
<td>1678</td>
<td>5018</td>
<td>0</td>
<td>0.33</td>
<td>III</td>
<td>10000</td>
<td></td>
</tr>
<tr>
<td>0.001</td>
<td>1</td>
<td>0.67</td>
<td>1244</td>
<td>3249</td>
<td>0.012</td>
<td>1.2</td>
<td>II</td>
<td>1000</td>
<td></td>
</tr>
<tr>
<td>0.005</td>
<td>1</td>
<td>0.67</td>
<td>446</td>
<td>1158</td>
<td>0.033</td>
<td>1.3</td>
<td>II</td>
<td>200</td>
<td></td>
</tr>
<tr>
<td>0.01</td>
<td>1</td>
<td>0.67</td>
<td>281</td>
<td>782</td>
<td>0.075</td>
<td>2.0</td>
<td>II</td>
<td>100</td>
<td></td>
</tr>
<tr>
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<td>0.67</td>
<td>36</td>
<td>101</td>
<td>1</td>
<td>3.4</td>
<td>I</td>
<td>10</td>
<td></td>
</tr>
</tbody>
</table>

¹Coefficient of bedrock erodibility.
‡Landslide threshold slope.
†Uplift rate.
§Coefficients of bedrock erodibility.

Figure 3. Graph indicating the resultant topography type (see text) for a range of U and K values. All model runs (n = 172) are included on this figure, and no overlap between topography types was detected for pairings of U and K, even with variation in S_c. Dotted lines indicate approximate transition between topography types.

Figure 4 shows how variation in parameters K and U affected synthetic topography for selected model runs that had the same threshold slope (20°) (the full data set from which this figure was derived can be found in the supplemental material, see footnote 1). From these plots, we concluded that drainage density and mean elevation are the most useful morphometric indicators for model comparison. These indicators show monotonic change with change in model parameters U and K, whereas steady-state hypsometric integral varied little except in cases of extremely low relief, and ruggedness number depended on two indicators (drainage density and mean elevation) that tended to have an inverse relationship, leading to complex and non-monotonic variation with changes in U and K. Relief changed in the same direction as mean elevation; however, mean elevation was less sensitive to local irregularities in high-elevation portions of the topography. The value of K affected drainage density, basin relief, and mean elevation most significantly. Since K is the primary control on the erodibility of a landscape, higher values of K (weaker rocks) led to limited relief development. For our highest K value (10 k.y.⁻¹), topographic relief over a 15 km² grid was less than 100 m; and when keeping all other parameters equal, an order-of-magnitude increase in K limited relief by as much as a factor of five. This relief limitation is related to the effect K has on drainage density. As K increases, drainage density increases as well, and by having closely spaced highly erosive channels, bedrock ridges cannot generate substantial relief between channels.

Rock uplift rate also had a strong control on steady-state topography, and acted to counter K. Higher uplift rates led to increased relief development and lower drainage density when all other parameters were held constant, which is consistent with the findings of Howard (1997). An order-of-magnitude increase in uplift rate commonly led to a three- to fourfold increase in model topographic relief in our model runs.

Figure 5 illustrates results from representative model runs, highlighting the effect of the composite parameter U/K and S_c on model topography. U/K is the primary control on relief production, and increased with either high uplift rates or with increasing rock strength (smaller K), or a combination of the two. Synthetic landscapes formed by model runs with high U/K values tended to be higher and dominated by mass-wasting processes rather than by fluvial processes. For model runs with U/K values of less than 100 m.k.y./k.y., relief was limited to less than 300 m over a 15 km² grid. On the other end of the spectrum, model runs with U/K values more than 1000 m.k.y./k.y. all had topographic relief exceeding 2500 m. Variation in model parameter S_c also affected model topography. This can also be seen in Figure 5. Increasing the threshold slope led to increased drainage density within type II topography by allowing for more closely spaced drainage divides and narrower canyons. Though drainage density was enhanced by higher values of threshold slope, and higher drainage density tended to correlate with lower mean elevation, we still produced higher topographic relief and higher mean elevation in our synthetic topography with higher threshold slopes for the same value of K. Because relief and grid lengths divided by total model pixel area; the absolute value of this maximum drainage density is sensitive to model pixel geometry), and type III topography had drainage density equal to zero. Our type II topographies had drainage density values ranging from just above zero to just below 0.0333 m⁻¹. These values can be normalized to the maximum drainage density for ease of comparison.

The size of the model grid affects the magnitude of most output morphologic characteristics (e.g., topographic relief increases with a larger grid size). Our model results are scalable, however, and the relative value of each of the morphologic indicators remains consistent, regardless of actual grid size. For model comparison to actual topography, we ran targeted model runs on grids that had the same area as our subsampled real DEMs in order to eliminate any ambiguity created by choice of grid size.

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Drainage density are both positively correlated with $S_c$, the ruggedness number showed an extremely strong positive correlation with $S_c$. The magnitude of the effect on topography from variation in $S_c$ was not nearly as strong as that of $U$ or $K$, since it mostly controlled the geometry of the interchannel divides—higher threshold slopes led to narrower, taller ridges.

For example, model runs with $U = 0.1$ m/k.y. and $K = 0.01$ k.y.$^{-1}$ on a 15 km$^2$ grid showed an increase in mean elevation from ~250 m to ~300 m with an increase in $S_c$ from 20° to 40°, while drainage density increased from 0.0017 to 0.0036 m$^{-1}$.

**Model Calibration: Geological Constraints**

In order to calibrate any model to a specific study site, an understanding of geological reality is necessary. In our study area in southern California, this required determining the ages of key sedimentary deposits. We discuss the techniques, data, and interpretation in some detail in the next section. We built upon recent detailed mapping by USGS personnel in the study area (Minor, 2004; Kellogg and Miggins, 2002; Stone and Cossette, 2000; Minor, 1999). Figure 6 shows the tectonic setting of our field area.

The upper reaches of the Cuyama River and its tributaries form dendritic drainage networks in several different lithologic units (Fig. 7). This fluvial incision postdates cessation of deposition in the ancestral Cuyama depositional basin, which occurred sometime since ca. 1 Ma, as evidenced by what is likely Bishop ash (760 ka) or possibly Glass Mountain ash (1.0–1.1 Ma) (Stone and Cossette, 2000) in the uppermost part of the Cuyama basin section. Since cessation of widespread basin filling, compressional deformation has occurred in the study area, which was likely caused by increasing transpression along the Big Bend section of the San Andreas Fault....
Figure 5. Model run results. All model runs are 15 × 15 km, with 30 m pixel size. (A) Relationships between $U/K$, $S_c$, and normalized drainage density. $U/K$ has dimensions of length if $m = 0.5$ and $n = 1$. Within the narrow range of $U/K$ values that result in type II topography (normalized drainage density between 0 and 1), higher threshold slope angle leads to an increase in drainage density by allowing for closer spacing of fluvial valleys, but drainage density is much more sensitive to variation in $U/K$ than to variation in $S_c$. (B) Relationships between $U/K$, $S_c$, and mean elevation.

Figure 6. Tectonic setting of study area in the Western Transverse Ranges, California, USA. Cuyama badlands are eroding late Cenozoic sedimentary units in the study area.
DeLong et al.

fault to the northeast of the study area. Deformation and erosion led to formation of an undulating planation surface that truncates bedding on the top of the basin-fill sediments. This surface was subsequently overlain by a widespread, but possibly discontinuous blanket of alluvium (Davis, 1983). Initiation of incision of the upper Cuyama River drainage basin has eroded the alluvium and underlying Cuyama Basin fills and left alluvium as ridge-top remnants and, in its largest preserved form, as the alluvium of the San Emigdio Mesa in the headwater reaches of the Cuyama drainage network. The age of this alluviation should therefore serve as the maximum age of the development of the modern drainage network in our study area.

We attempted to calibrate the model using our understanding of late Pleistocene landscape development in three lithologic units: (1) The Pliocene-Pleistocene Morales Formation, a weakly to moderately consolidated conglomerate and coarse sandstone unit that was deposited in a terrestrial environment in response to late Cenozoic compressional tectonics (Page et al., 1998); (2) the Pliocene Quatal Formation, a generally silty to clayey terrestrial sandstone with interbedded conglomerates; and (3) the Eocene Matilija sandstone, a complex marine unit that has facies of sandstone, conglomerate, siltstones, and shale (Minor, 2004). Each of these units is weak enough to have experienced significant drainage development during the late Quaternary; however, enough differences exist in characteristics such as sedimentary structure, bedding attitude, and material properties such that each formed somewhat different landscape morphologies in response to similar forcing.

The most dramatic lithologic contrasts exist across the Big Pine fault, a northeast-trending oblique reverse fault that places Eocene rocks against Miocene through Pleistocene rocks (Minor, 2004, 1999). While this created a favorable configuration of differing rock types within a single drainage basin, possible late Pleistocene vertical movement on the Big Pine fault complicated our analysis somewhat by introducing possible variation in $U$ across the Big Pine fault. We addressed this by analyzing spatial relationships between equal-age alluvial deposits across the fault zone.

While our study area is referred to as the Cuyama badlands, it should be noted that much of the area does not display the classic features of badland topography. Topographic development in the three lithologic units that we studied was not characterized by densely gullied, fluvially dominated slopes, but was rather made up of generally planar threshold slopes and fluvial channels. It appears that coupled shallow landsliding and fluvial incision have been the dominant late Quaternary surface processes. We found no evidence for debris-flow incision, and soil production is limited to several centimeters on vegetated slopes, which form a mosaic with more recently failed threshold slopes. Deep soils are limited to interchannel areas on which San Emigdio Mesa equivalent alluvium is preserved. This limited scope of active processes, the moderate relief of tens to hundreds of meters, and
the moderate regional uplift rates, we believe, justify our choice of a very simple landscape development model.

**Surface Exposure Dating, Burial Dating, Uplift Rate, and Model Calibration**

The alluvium that forms San Emigdio Mesa was deposited discontinuously over an undulating Pleistocene erosion surface (Davis, 1983). We correlated this alluvium with alluvial remnants preserved throughout the upper Cuyama drainage basin on both sides of the Big Pine fault. These correlations were based on map relationships, similar and indicative crystalline source material, soil color, and soil texture. Soils formed on all correlative deposits tended to have 5YR color, incipient argillic or cambic B-horizon formation, and no visible petrocalcic accumulation. These alluvial deposits form a measurable datum that marks the most recent time predating fluvial network incision. We therefore use the age of the San Emigdio Mesa as an estimate for the age of the erosional landscape in the upper Cuyama region. Previous workers have suggested that the alluvium of San Emigdio Mesa (the large preserved alluvial remnant that predates formation of the drainage network) (Fig. 7) is as young as late Pleistocene in age (Dibblee, 1972; Davis, 1983), though their correlations with dated sediments elsewhere were not justified in detail.

**10Be Surface Exposure Dating**

The 10Be surface exposure dating method allows us to measure the time of exposure of a geomorphic surface by taking inventory of in situ–produced radionuclides that accumulate in the upper 1 m or so of the surface as a product of cosmic ray bombardment. For 10Be surface exposure dating, we sampled material from three flat-topped granitic boulders partially embedded in the alluvial surface of San Emigdio Mesa in order to approximate the age of the underlying alluvial deposit (Fig. 8A). These were selected with the criteria of showing no obvious signs of spallation, weathering, or past burial and excavation (Fig. 8B). Soils developed on this alluvial surface were suggestive of long-term stability, and the lack of gullyling or removal of upper fine-grained and reddened soil horizons were our criteria for lack of past burial and excavation. Isotopic analysis of 10Be abundance in quartz was carried out at the Purdue University Rare Isotope Measurement Laboratory according to standard procedures. These data were corrected for sample thickness and topographic shielding, and were then corrected for latitude, longitude, elevation, and past geomagnetic effects following Pigati and Lifton (2004). In order to use the most accurate cosmogenic production rate for 10Be, we also corrected the raw data of Stone et al.'s (1998) Younger Dryas–aged samples from Scotland. From this we took the long-term integrated high-latitude sea-level 10Be production rate to be 4.35 atoms g−1 yr−1. Using Clark et al.'s (1995) data, we followed the same procedure to calculate an integrated production rate of 4.58 atoms g−1 yr−1, but opted to use Stone's data because we judged it to have better independent age control. Following Partridge et al. (2003), we used a mean life of 10Be of 1.93 ± 0.10 Ma; for discussion of ambiguity related to this value, see note 34 therein.

Our 10Be results (Table 2) indicate a late Pleistocene age for the San Emigdio Mesa. Sample B has a 10Be age of only 14.3 ka, which records the most recent localized deposition on the surface of the San Emigdio Mesa. Sample B is from an area that is below a set of steep, small drainages that are graded to the surface of the mesa. These drainages appear to be relatively inactive currently, and appear to have last deposited material onto the mesa during the latest Pleistocene. The other two boulders sampled for cosmogenic 10Be dating are from areas that did not receive depositional erosion as recently (their exposure ages were 32.3 ± 1.0 and 28.2 ± 0.9), and likely reflect the most recent widespread deposition on the alluvial surface that predated the most proximal incision. The relative proximity of the sample locations on the alluvial deposit to the sediment source area, the degree of soil formation, and the youth of all exposure ages compared to previous estimates and our optically stimulated luminescence (OSL) ages (see next section) lead us to believe that inherited cosmogenic radionuclides from predepositional exposure are negligible. Though we are unable to quantify it, any post-depositional erosion of the sampled boulders would lead to systematic underestimation of their exposure age. Our 10Be data does show scatter, and this and the small number of samples necessitates the use of OSL burial dating to further support our interpretations.

**OSL Burial Dating**

Optically stimulated luminescence (OSL) dating allows us to determine the time elapsed since burial of sediments. It assumes that electrons stored in crystal imperfections are removed by light exposure during transport, and they reaccumulate at a measurable rate after subsequent burial. We employed OSL on three samples of bedded alluvial sands from near the top of an ~60 m slope of nearly horizontally bedded sandy alluvium that unconformably caps deformed basin fill on the San Emigdio Mesa. Bedded sands were sampled below the main pedogenic zone in order to assure a sample of unmixed primary sediment (Fig. 8C). These were sampled using opaque ABS pipe without exposing the sediment to light during sampling. Dose rate measurements were made directly in the field using a portable 4-channel gamma spectrometer calibrated by personnel at the USGS. Environmental dose rate values were calculated using the conversion factors of Adamicz and Aitken (1998) and the grain-size attenuation factors outlined in Aitken (1985). Present-day water content values were assumed to be representative of those pertaining to the full burial period and were assigned relative uncertainties of ±50%.

Equivalent dose (De) analysis was undertaken at the Oxford University Luminescence Research Group Laboratories. Coarse-grained (125–180 μm) quartz De measurements were made using the single aliquot regeneration (SAR) protocol developed by Murray and Wintle (2000). Individual De estimates were calculated for 15 aliquots (composed of 100–300 grains) from each of the three samples. Sample bleaching characteristics were then assessed from these populations of individual De estimates using the approach suggested recently by Bailey and Arnold (2006). All three of these samples appeared to have been adequately bleached prior to deposition and burial following this analysis. Final burial dose estimates were therefore calculated using the “central age model” of Galbraith et al. (1999).

Results from the three OSL samples from San Emigdio Mesa (Table 3) indicate a late Pleistocene depositional age for bedded sands in the San Emigdio Mesa alluvial deposit. These dates are noticeably older than the cosmogenic ages from boulders on the surface of the deposit; however, evaluation of the context of each dating method suggests that they give complementary results. Boulders on the surface of an alluvial deposit could have been deposited more recently than bedded sands exposed by fluvial incision that are below the primary soil-formation horizon (Fig. 8D). Our data suggest that at ca. 60 ka, bedded fluvial sands were being deposited on an unincised San Emigdio Mesa, whereas by ca. 30 ka, deposition of coarse material was occurring in a more proximal setting, and incision of the mesa was possibly occurring in more distal regions.

**Application of Geochronology to Model Calibration**

Since we assumed that our model begins with the earliest incision into an undissected surface, it was necessary to constrain the timing of initiation of drainage network incision into the upper Cuyama basin in order to compare the basin topography with model results. It is possible that
DeLong et al.

14.3 ± 0.7
28.2 ± 0.9
32.3 ± 1.0
60.64 ± 6.41
57.55 ± 6.52
69.53 ± 5.74

OSL sample
10Be sample

Drainages Graded
to San Emigdio
Mesa surface

Youngest alluvial
deposit

Granitic Mountains

Y oungest alluvial
deposit

Figure 8. San Emigdio Mesa and geochronological sample sites. (A) Aerial photograph of part of San Emigdio Mesa. Geochronological sample locations are marked with dates (see Tables 2 and 3 for complete data), and generalized surficial geology shows two distinct areas of late Quaternary deposition as constrained by direct field observation (see Fig. 6 for location map). (B) Boulder sampled for 10Be surface exposure dating from San Emigdio Mesa. Global positioning system (GPS) unit on boulder surface is 17 cm long. (C) Optically stimulated luminescence (OSL) sample location; notice faint fluvial bedding and primary pedogenic zone well above sample location. Gamma spectrometer probe is ~12 cm in diameter. (D) Setting of typical OSL sample. Alluvium unconformably overlies deformed sedimentary bedrock below. Underlying badland-forming bedrock in this photo is local exposure of Caliente Formation, which was not analyzed in this study due to limited extent in study area.

TABLE 2. COSMOGENIC 10Be DATA AND RESULTS

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Sample description</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Elevation (m)</th>
<th>S3</th>
<th>Thickness correction</th>
<th>10Be† (atoms g–1)</th>
<th>10Be age‡ (ka)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>Granitic boulder</td>
<td>34°48′22″N</td>
<td>119°15′36″W</td>
<td>1575</td>
<td>1.0001</td>
<td>1.027</td>
<td>436,000 ± 14,600</td>
<td>32.3 ±1.0</td>
</tr>
<tr>
<td>B</td>
<td>Granitic boulder</td>
<td>34°49′4″N</td>
<td>119°15′30″W</td>
<td>1580</td>
<td>1.0001</td>
<td>1.044</td>
<td>179,100 ± 10,400</td>
<td>14.3 ±0.7</td>
</tr>
<tr>
<td>C</td>
<td>Granitic boulder</td>
<td>34°48′53″N</td>
<td>119°15′5″W</td>
<td>1627</td>
<td>1.0001</td>
<td>1.027</td>
<td>390,600 ± 13,900</td>
<td>28.2 ±0.9</td>
</tr>
</tbody>
</table>

†Topographic shielding factor.
‡Confirmed for chemical blank value, sample thickness, and topographic shielding. Error displayed is analytical error only.
§Age determined following Pigati and Lifton (2004) with a time-integrated high latitude, sea level 10Be production rate of 4.35 atoms yr–1. No erosion or burial assumed.
the initiation of incision predated our cosmogenic ages, since they date deposition in the upper parts of the San Emigdio Mesa alluvial deposit. Our OSL dates were from a time when deposition was still occurring on San Emigdio Mesa, and likely throughout much of our study area. We suggest initiation of basin incision occurred during the last 60 k.y., but no more recently than approximately the last 30 k.y. Therefore, for model-run comparisons, we expect dynamic steady state to be approached in ~60 k.y., especially in more distal parts of the study area.

In addition to model run time, a key parameter needed for our model calibration was $U$, which is the vertical uplift rate of the landscape. In the model, this is a rock uplift rate; the actual elevation change of surfaces may vary due to local erosion. In Cuyama Valley, we assumed that this value was a combination of the direct tectonic uplift rate of the region and the local base-level lowering rate transmitted through the landscape from the main-stem Cuyama River. Since the fluvial system occupies a characteristic position ~60–80 m below ridge tops and preserved alluvial caps north of the Big Pine fault, we took the uplift rate to be on the order of 1–2 m/k.y. This equates rock uplift rate, surface uplift rate, and the local fluvial incision rate if we assume the alluvium-capped ridges were deposited at local base level and subsequently uplifted with minimal erosion.

On the south side of the Big Pine fault, which was likely thrusting northward during the late Pleistocene, relief is up to 80–100 m below alluvium that we interpreted as equal-aged to that across the fault and to the alluvium of San Emigdio Mesa. This suggests a higher value for $U$ of 1–3 m/k.y. based on our interpreted age of the ridge-capping alluvium. These values are comparable to other regional estimates for tectonic uplift in coastal and near-coastal California (Ducea, et al., 2003; Page et al., 1998; White, 1992).

**DISCUSSION**

The relative simplicity of our model allows for a detailed analysis of each of the model parameter’s effects on synthetic landscape development. Many previous studies have demonstrated the stream power law’s ability to simulate bedrock channel development (e.g., Stock and Montgomery, 1999, and references therein). By coupling this tested method to simulate channel development with a simple yet realistic process model for hillslope development in tectonically active terrains, we believe our model appropriately simulates the primary controls on landscape development over a variety of landscape process rates and spatial scales. Furthermore, we believe our model results can be compared to actual landscapes in an effort to calibrate model parameters, and the response of synthetic landscapes to model behavior is likely mirrored to a reasonable extent in actual landscapes.

**Values of $K$ and Relationship between $K$ and $U$**

Variation in model parameter $K$ had a profound effect on model topography. By analogy, we expect rock resistance in tectonically active areas to be a primary control on landscape morphology. Our model results indicate that $K$ can only vary over less than three orders of magnitude in landslide-dominated tectonically active areas (as defined by model-derived type II topographies) (Fig. 3). The actual range of $K$ values for various rock types in uplifting areas appears to correlate positively with $U$. This is perhaps consistent with the observations of Snyder et al. (2000) that $K$ undergoes a “dynamic adjustment” to tectonic changes, though in our model $K$ strictly controls the rock resistance to erosion at a given point regardless of uplift rate or erosive process, and does not adjust to $U$. Our model does not address dynamics thought to be related to change in erosional process, such as sediment flux change, channel geometry change caused by increased debris flows, or increased orographic precipitation, which Snyder et al. (2000) suggested occurs with tectonic perturbation. It does seem likely, however, that we might expect a change in effective $K$ with increased uplift rate, since in nature the erodibility of bedrock is related to the sediment flux over the water-rock interface (Gilbert, 1877; Sklar and Dietrich, 2001), which is in turn a function of $U$ in orogenic settings.

In general, we found linear relationships between the upper and lower possibilities for $K$ that form type II topography. Upper values for $K$ (weak rock leading to lower-relief topography) in type II topography in our data set had the following relationship with $U$:

$$K = 0.02U - 0.004. \quad (2)$$

Lower values for $K$ (resistant rock leading to higher-relief topography) in type II topography in our data set had the following relationship with $U$:

$$K = 0.002U - 0.0004. \quad (3)$$

Equations 2 and 3 do not strictly bind the range of parameters that form type II topography, but they show the range of parameters that do form type II topography in our order-of-magnitude evaluation of landscape response to variations in $K$. Figure 9 shows the range of likely $K$ values for a range of values of $U$. It also shows that data from Snyder et al. (2000), which are a number of estimates for $K$ over a range of uplift rates from coastal northern California, fell within our envelope that defines the relationship between $U$ and $K$ in type II topography. Furthermore, we found that all our type II model results have $U/K$ values between 50 and 1000 m·k.y. /k.y.; all type I model results have $U/K$ values of 10 m·k.y./k.y. or lower; all type III model results have $U/K$ values of 5000 m·k.y./k.y. and above; and there is no overlapping of topography type for each unique $U/K$ value. We therefore conclude that in most tectonically active terrains, values for $U/K$ should fall above 10 and below 5000 m·k.y./k.y.

These relationships allow us to better constrain the range of possible values of $K$ in landscapes that have known uplift rates.

When the results of Stock and Montgomery (1999) are evaluated in terms of these constraints on the relationships between $K$ and $U$, it may be possible to better understand the wide range of variation in their calculated values of $K$, which range from $10^{-7}$ m$^2$/yr$^{-1}$ to $10^{-3}$ m$^2$/yr$^{-1}$. The lowest values they calculated for $K$ come from tectonically quiescent and resistant crystalline rocks of Australia, the intermediate values for $K$ come from volcanic rocks from both tectonically active California and also Hawaii, where tectonic uplift is supplanted by addition of rock mass from lava flows, so $U$ is essentially a fairly high base-level lowering rate. Their highest values for $K$ come from Japan, where presumably weak mudstones are evaluated for $K$ over a very short (5 k.y.) period in response to a migrating knickpoint, which seems likely to be approximated by a high value for $U$. Their data reveal that in order to have a wide range of

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Depth (m)</th>
<th>Dose rate (Gy/k·y.)</th>
<th>$D$ (Gy)</th>
<th>Burial age (ka)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SEM01</td>
<td>5</td>
<td>4.07 ± 0.23</td>
<td>246.6 ± 22.1</td>
<td>60.64 ± 6.41</td>
</tr>
<tr>
<td>SEM02</td>
<td>6</td>
<td>4.25 ± 0.23</td>
<td>244.8 ± 24.3</td>
<td>57.55 ± 6.52</td>
</tr>
<tr>
<td>SEM03</td>
<td>9</td>
<td>4.14 ± 0.23</td>
<td>288.1 ± 17.3</td>
<td>69.53 ± 5.74</td>
</tr>
</tbody>
</table>

OSL—Optically stimulated luminescence; Gy—unit of absorbed dose.
values for $K$, an analysis must be done of several basins with both a wide range of rock types and a wide range of uplift rates.

**Model Calibration Using Real Topography**

Our efforts to calibrate the stream power law using this model provide estimates for $K$ from the upper Cuyama region. Table 4 shows the values we measured in the field and extracted from DEMs for our three target lithologies for comparison to controlled model runs. For this part of our study, we ran targeted model runs on grids with the same area as our subsampled DEM grids (see Fig. 7) for each lithologic unit. Running targeted model runs required estimation of model parameters $m$ and $n$. For these, we relied on analysis of DEM-derived slope and area data. Following Snyder et al. (2000), in the case of a steady-state landscape in which uptake rate $U$ and coefficient of erosion $K$ are uniform along the channel reach, Equation 1 can be solved for equilibrium slope ($S_e$):

$$S_e = kA^n$$  \hspace{1cm} (4)

where

$$k = (U/K)^{1/n}$$  \hspace{1cm} (5)

is the steepness index and

$$\theta = m/n$$  \hspace{1cm} (6)

is the concavity index. $\theta$ and $k$, in Equation 4 can be calculated using the regression of channel-slope and drainage-area data. Rearranging Equation 4 to the form of the equation of a linear regression of the logarithm of slope and area data gives:

$$\log S = -\theta \log A + \log k,$$  \hspace{1cm} (7)

in which $S$ is local channel slope. Therefore $mn$ is approximated by the slope of the regression line on a log plot of slope-area data, and the value of $(U/K)^{1/n}$ is the y-intercept of the regression (Snyder et al., 2000).

We used DEM-derived slope and area data in order to estimate $mn$ for each lithologic unit. We conservatively confined our analysis to channel reaches that had drainage areas greater than $10^2 \text{ m}^2$, to assure that our analysis only included zones of fluvial incision and not headwater mass-movement or other geomorphic processes. Figure 10 shows our results visually. We extracted the data from the logarithm of slope of Strahler stream segments (as opposed to basin pixels or channel pixels, in an effort to reduce scatter in DEM data) versus drainage area. Data scatter was evident and $R^2$ values ranged from 0.27 to 0.42. Our methods differed from Snyder et al. (2000) in how we handled the data scatter from low-order channels. They used a map-based data extraction in which each 10 m elevation range represented a single data point along a single main channel, whereas we included all tributaries above our drainage-area threshold; this introduced scatter that served to lower our $R^2$ values, but it captured variability in tributary channel slopes that we felt was important in a 3D model calibration. Snyder et al. (2000) performed several methods of data extraction and analysis to generate estimates for $mn$, and found that field, DEM, and manual estimation from topographic maps provided similar and satisfactory estimates.

We were able to match morphologic indicators from the upper Cuyama region with model run outputs with a good degree of success. Figure 11 shows what we believe to be our most useful comparative morphologic parameters—mean elevation and drainage density, plotted against key model parameters—and also our values for the three lithologic units from the upper Cuyama region. Mean elevation and drainage density were highlighted because they were the morphologic indicators that were most sensitive to model parameter changes. We subsampled USGS DEMs in rectangular grids that contained the largest possible areas of a single rock type for topographic characterization.

Using Figure 11, we were able to constrain the values for $K$ for each of the lithologic units. For the Matilija Formation, we suggest an appropriate $K$ value is on the order of $0.09–0.25 \text{ m}^3/\text{k.y.}^{-1}$; for the Quatal Formation, $K$ appears to be between $0.1$ and $0.3 \text{ m}^3/\text{k.y.}^{-1}$; and the Morales Formation appears to have a $K$ value of $0.15–0.3 \text{ m}^3/\text{k.y.}^{-1}$. These values are on the high end of the range proposed by Stock and Montgomery.
(1999), which is not surprising given that they are fairly poorly indurated sedimentary units that should be among the “weakest” rocks able to support moderate-relief topography.

The techniques used by Snyder et al. (2000) also allowed for an independent estimate of $U/K$ based on the channel steepness index. Using Equation 7, our slope and area data, and our estimates for uplift rate $U$ from geologic data, we obtained estimates for $K$ that are similar, but generally higher than our estimates based on matching morphometric indicators from targeted model runs. Using the slope and area method, $K$ for the Morales Formation is between 0.4 and 0.8 m$^0.2$ k.y.$^{-1}$, for the Quatal Formation, it is 0.4–0.6 m$^0.4$ k.y.$^{-1}$, and for the Matilija Formation, $K$ is between 0.13 and 0.25 m$^0.4$ k.y.$^{-1}$. The slope and area method for calibration of $K$ seems to work well for bedrock channel reaches, though calibrating the stream power law with only a portion of the landscape makes it susceptible to error in extraction of slope and area values from DEMs. Given that the $R^2$ values for our slope and area were fairly low, we placed more confidence in the use of our full hillslope-channel landscape development model calibration techniques. The comparative morphologic indicators we used for model calibration were simple to extract from DEMs with a minimum of error (especially mean elevation), and this method allowed us to integrate the entire landscape into model calibration.

A possible complication in our comparison of real landscape development to synthetic landscape development is proper temporal scaling. As discussed previously, we believe our models should produce comparative topography with a model run time of ~60 k.y. We approximated model run time from geologic data; however, model runs reached a “dynamic steady state” in which topographic configuration was no longer highly dependent on model run time (Willett, 1999), so the age of the topography need only serve as a minimum value for model run time in steady-state regimes. In order to address the possibility that the comparison landscapes in California are not appropriately approximated by steady-state model runs, we also extracted morphologic indicators as a function of time for an evolving landscape for a set of model parameters to determine how these indicators behaved before steady state was achieved.

Our analyses of the Quatal and Matilija Formations appear to be robust, and the modeled processes achieved steady state in accordance with our geochronological constraints. Our model runs designed to target landscape development within the Morales Formation, however, had difficulty reaching a fully dissected form by 60 k.y. In addition, the actual value for hypsometric integral $m/n = 0.39$

\[ R^2 = 0.40 \]

Along-channel slope

\[ m/n = 0.29 \]

\[ R^2 = 0.42 \]

Drainage area (km$^2$)

\[ m/n = 0.37 \]

\[ R^2 = 0.27 \]

\[ m/n = 0.29 \]

\[ R^2 = 0.42 \]

\[ m/n = 0.39 \]

\[ R^2 = 0.40 \]

Figure 10. Slope and drainage area data extracted from 30 m U.S. Geological Survey digital elevation model (DEM) data. Each data point represents a single Strahler stream segment in order to avoid slope errors near “stair-steps” caused by DEM production from topographic maps. Vertical dashed line indicates lower drainage area bound for analysis domain. Slope of gray line indicates the negative of $m/n$. 

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from the Morales Formation was 0.5, which was likely an indicator that this landscape had not quite reached a steady-state form.

In order to evaluate topographic sensitivity to landscape development during the time before dynamic steady state was reached, we ran targeted model runs for a range of model run times approaching the time it takes to achieve steady state. All of our steady-state model runs landscapes of type II and type III achieved a hypsometric integral of ~0.32–0.40, and the time evolution of the hypsometric integral, mean elevation, topographic relief, and drainage density can be seen in Figure 12. These values indicate that it is possible to make appropriate estimations for \( K \) in landscapes that are approaching, but have not yet achieved steady state. Based on this, we believe that our estimated value for \( K \) for the Morales Formation is a bit too low, and that the portion of the landscape that we subsampled in the Morales Formation has not reached its ideal form. Figure 13 shows model run output topography as it approaches steady-state configuration. Noting that hypsometric integral and mean elevation lowers as steady state is approached, and drainage density increases, we propose a better approximation for \( K \) for the Morales Formation of ~0.2–0.4 \( m^{0.4}/k.y. \). This argument is supported by the observation that the San Emigdio Mesa is a preserved surface that has not yet been significantly incised, and the Morales Formation is the most proximal bedrock unit, so it may have been subject to erosion for less time than the more distal Quatal and Matilija Formations, and therefore it has not yet achieved steady state. This basin-scale time-transgressive nature of temporal development of hypsometric integral is illustrated in Figure 14.

While our estimates for the values of \( K \) for the three lithologic units are very similar, our model parameterization using geologic data introduced some interesting complexity that must be taken into account when comparing \( K \) values to rock properties. First, our estimation of values for the area exponent \( m/n \) from topographic data affected the erosivity of the model. Our estimate of \( m = 0.3 \) for the Morales Formation limited the model’s erosiveness for a given \( K \) value, so while our estimate for \( K \) for the Morales Formation was similar to the others when our estimate for \( m \) is taken into account. Secondly, our geologically derived estimates for differing uplift rate across the Big Pine fault affected the values for \( K \) across the fault. The Eocene Matilija Formation appears to be a more indurated, less erodible lithology; however, the coupling of \( U \) and \( K \) (Snyder et al., 2000) leads to an estimate for \( K \) for the Matilija Formation that is only marginally lower than those for the less-indurated Quatal and Matilija Formations, though the rocks seem to differ more than is suggested by the \( K \) values upon field inspection. Finally, the differing landslide threshold slopes affect model topography and lead to differing mean elevations and drainage densities for similar \( K \) values. In particular the higher threshold slopes in the Quatal and Matilija Formations have the effect of increasing drainage density and mean elevation for a similar value of \( K \).

These observations imply that while \( K \) may be somewhat effective at defining resistance to fluvial erosion, when using a stream power law–based model to model landscape development, \( K \) should not be thought of as singularly reflecting material properties of a given

Figure 11. Calibration of \( K \) from key morphologic parameters of our three study lithologies. Each rock type has a curve corresponding to high and low estimates for \( U \) from geologic data, and the horizontal bars are the actual values extracted from digital elevation models (DEM) for each lithology matched to the model curves. Shaded vertical bars indicate the range of likely values for \( K \) for each rock type. Morales Formation model runs were done on a 603 × 603 cell grid with 10 m grid cell size, \( m = 0.3 \), and \( n = 1 \). Quatal Formation model runs were done on a 309 × 309 cell grid with 10 m grid cell size, \( m = 0.4 \), and \( n = 1 \). Matilija Formation model runs were done on a 131 × 131 cell grid with 10 m grid cell size, \( m = 0.4 \), and \( n = 1 \). These grid geometries were chosen to match the subsampled real DEMs.
lithology. Material properties are also reflected in the parameters dictated by landslide threshold slope and channel concavity (m/in). Uplift rate also affects topography independently of the material properties of the bedrock. For example, though the Quatal and Matilija Formations appear to have similar \( K \) values and threshold slopes, the Matilija holds up a higher mean elevation and relief, presumably owing to the higher value for uplift rate that we estimated from geologic data. Also, though the Morales and Quatal Formations also appear to have a fairly similar \( K \) value, they have much different drainage density, which is likely related to the higher threshold slope value of the Quatal Formation, presumably caused by the finer-grained cohesive nature of the Quatal Formation.

These observations lead us to suggest that no singular “rock-strength” parameter (e.g., Selby, 1980) is available to characterize landscape-scale erosional behavior of a rock type, and a more holistic approach that integrates overall morphology and uplift rate is necessary in order to relate topography to lithology and vice versa. This exercise shows that not only is the stream power law extremely sensitive to changes in the value for \( K \), but in order to calibrate stream power law–based values of \( K \), one needs to take into account all other model parameters. Furthermore, an estimated value of \( K \) extracted using the techniques proposed in this study does not directly correlate to rock erodibility, because channel profile shape and uplift (or base-level lowering rate) all affect modeled erosion. Based on these results, we propose that drainage density and mean elevation (or topographic relief) can be used to uniquely infer \( K \) to within a relatively narrow range in tectonically active landscapes with steady-state configurations if other model parameters are also carefully calibrated.

CONCLUSIONS

Three-dimensional modeling that utilizes the stream power law for fluvial erosion and threshold landsliding for hillslope development allows us to analyze the effects of parameters such as uplift rate, bedrock erosivity, threshold slope, channel concavity, and time effect on landscape development. We suggest that by careful comparison of (1) actual landscape morphology via field and DEM analysis, and (2) actual landscape development process rates from geochronology to synthetic topography derived from a numerical model with carefully controlled parameters, we can calibrate modeling efforts, and in particular, narrow our range of estimates for \( K \). We suggest that characterization of m/n, landslide threshold slope, mean elevation, topographic relief, drainage density, and hypsometric integral are necessary for comparison of actual topography to synthetic topography. In our study area, three late Cenozoic sedimentary units were estimated to have \( K \) values on the order of 0.3–0.09 m\(^{0.2–0.4}\) k.y.\(^{-1}\). We addressed possible complications from temporal and spatial scaling, and suggest that even complex and/or non-steady-state real topography can be compared to idealized synthetic topography with some measure of success. Work on widely different rock types and spatial scales will be necessary to further validate our results.

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Figure 13. Oblique shaded-relief images of the development of synthetic topography through time. Model parameters are given in Figure 10 caption.

Hypsometric Integral = 0.40  Hypsometric Integral = 0.53

Figure 14. Oblique shaded-relief image of synthetic topography that has not reached steady state and a subsample of that topography, illustrating that the hypsometric integral (the primary indicator of steady-state topography in our study) approaches that of steady state away from a preserved plateau. Terminal steady-state hypsometric integral for this model run is ~0.35.