Numerical modeling of the Cenozoic geomorphic evolution of the southern Sierra Nevada, California

Jon D. Pelletier

Department of Geosciences, University of Arizona, Gould-Simpson Building, 1040 East Fourth Street, Tucson, Arizona 85721-0077, USA

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Abstract

Recent geomorphic studies suggest that significant (∼1.5 km) late Cenozoic surface uplift occurred in the southern Sierra Nevada, a conclusion that is difficult to reconcile with recent stable-isotopic paleoaltimetry studies. Numerical modeling can play an important role in resolving this dispute. In this paper I use two models of bedrock channel erosion, the stream-power model and a sediment-flux-driven model, to test hypotheses for the fluvial Cenozoic geomorphic evolution and surface uplift history of the southern Sierra Nevada. Cosmogenic data for upland erosion and river incision rates allow each model parameter to be uniquely constrained. Numerical experiments using the sediment-flux-driven model suggest that the modern southern Sierra Nevada was constructed from a 1.0-km pulse of range-wide surface uplift in the latest Cretaceous (∼60 Ma) and a 0.5-km pulse in the late Miocene (∼10 Ma). The persistent geomorphic response to latest Cretaceous uplift in this model is the result of limited “cutting tools” supplied from the upland low-relief Boreal Plateau. This uplift history correctly predicts the modern topography of the range, including the approximate elevations and extents of the Chagoopa and Boreal Plateaux and their associated river knickpoints. Numerical experiments using the stream-power model are most consistent with a 1-km pulse of uplift in the late Eocene (∼30 Ma) and a 0.5-km pulse in the late Miocene (∼7 Ma). Both models suggest that the remaining rock uplift required to produce the 4-km peak elevations of the modern southern Sierra Nevada was produced by flexural-isostatic uplift in response to river incision. The balance of evidence, including the dominance of sediment-flux-driven erosion in granitic rocks, previous paleoaltimetry studies, and the timing of sediment accumulation in the Great Valley, support the conclusions of the sediment-flux-driven model, i.e. that the Sierra Nevada experienced range-wide surface uplift events in the latest Cretaceous and late Miocene. More broadly, these results indicate that nonequilibrium landscapes can persist for long periods of geologic time, and hence low-relief upland landscapes do not necessarily indicate late Cenozoic surface uplift.

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1. Introduction

In the past few years, stable-isotopic and geomorphic approaches have been used to infer the Cenozoic surface uplift history of the Sierra Nevada of California (e.g. Poage and Chamberlain, 2002; Clark et al., 2005b; Mulch et al., 2006), the Himalaya and southern Tibet (e.g. Clark et al., 2005a, 2006; Currie et al., 2005; Grujic et al., 2006), and the central Andes (e.g. Ghosh et al., 2006; Barke and Lamb, 2006). In some cases, strong arguments exist both for and against late Cenozoic surface uplift for a particular region. In the Sierra
Nevada, for example, the presence of a slowly eroding plateau perched 1.5 km above narrow river canyons suggests late Cenozoic surface uplift of ~1.5 km in the southern part of that range (Clark et al., 2005b). Stable-isotope paleoaltimetry of Eocene gravels (Mulch et al., 2006), however, imply that the northern Sierra Nevada achieved an elevation of at least 2.2 km by Eocene time. These results are not necessarily contradictory since they represent different parts of the range, but they require distinctly different Cenozoic surface uplift histories in the northern and southern Sierra Nevada in order to be reconciled.

Three additional lines of geomorphic evidence suggest that significant late Cenozoic rock uplift occurred throughout the Sierra Nevada, bolstering the claim for widespread late Cenozoic surface uplift. Unrath (1991) documented 1.4° of post-late-Miocene westward tilting of the northern Sierra Nevada based on the stratigraphy of the Great Valley. Huber (1981, 1990) documented approximately 700 m of stream incision/rock uplift along the mainstem San Joaquin and Tuolumne drainages (southern and central Sierra Nevada) since 10 Ma. Wakabayashi and Sawyer (2001) extended Huber’s work to drainages in the northern Sierra Nevada, documenting as much as 1 km of late Cenozoic (<10 Ma) incision but limited Eocene–Miocene incision in drainage headwaters. These authors argued that stream incision and surface uplift are equivalent near the crest of the range because of the low cosmogenic erosion rates measured there (i.e., ~0.01 mm/yr). Stock et al. (2004, 2005) dated cave sediments cosmogenically to document a pulse of relatively high channel incision rates (~0.3 mm/yr) between 1.5 and 3 Ma in the South Fork Kings River and nearby drainages of the southern Sierra Nevada. The broad geographic range of these studies clearly indicates that significant late Cenozoic rock uplift occurred throughout the range. Similarity in the large-scale topographic form of the northern and southern Sierra Nevada (i.e. both have asymmetric profiles with steep eastern escarpments and gently dipping western slopes), as well as similar Cenozoic exhumation rates (Cecil et al., 2006), suggest that the northern and southern Sierra Nevada may have undergone similar surface uplift histories. Nevertheless, the northern and southern parts of the range have significant structural differences: the northern Sierra Nevada is a west-dipping tilt block, while deformation within the southern Sierra Nevada is accommodated on a series of normal faults (Clark et al., 2005b).

Despite extensive research, the relationship between rock uplift and surface uplift in the Sierra Nevada remains uncertain for two reasons. First, surface uplift triggers knickpoint retreat along mainsteam rivers, but the time lag between incision at the range front and incision tens of kilometers upstream is not well constrained. Thermochronologic data have been collected from many sites (e.g. House et al., 1997, 1998, 2001; Clark et al., 2005b; Cecil et al., 2006), but the westernmost samples are located approximately 30 km from the range front. In the southern part of the range, these data provide a 32 Ma maximum age for the onset of stream incision at sample localities. Range-wide surface uplift could have occurred significantly earlier than 32 Ma, however, given the time that may have been required for knickpoints initiated at the range front to propagate 30 km or more upstream. Knickpoint retreat rates of 1 m/kyr, for example, would result in a 30 Myr time lag between range-wide surface uplift and knickpoint passage at westernmost sample localities. Second, rock uplift, stream-incision, and local surface uplift can all occur in the absence of range-wide surface uplift. As stream incision removes topographic loads from the crust, for example, isostatic rebound raises slowly eroding upland plateau remnants to elevations much higher than the original, regionally extensive surface. As such, no simple relationship exists between the timing of local surface uplift and range-wide surface uplift.

The goal of this paper is to evaluate hypotheses for range-wide surface uplift of the southern Sierra Nevada and to refine our understanding of the spatial and temporal distribution of uplift and erosion using numerical landscape evolution modeling. This paper focuses on the southern Sierra Nevada because of the limited impact of Plio-Quaternary glaciation in valleys of this part of the range (Clark et al., 2005b). Cosmogenic erosion rate studies provide key constraints on numerical model parameters, enabling the relationship between specific uplift histories and the topographic evolution of the range to be uniquely determined, including the effects of transient knickpoint migration and flexural-isostatic response to erosion. In the next section I review key aspects of the geomorphology and geochronology of the southern Sierra Nevada that provide calibration or validation data for the numerical model.

2. Geomorphology and rates of landscape evolution in the Sierra Nevada

Two distinct topographic surfaces have long been recognized in the southern Sierra Nevada (Webb, 1946) (Fig. 1). The Boreal Plateau is a high-elevation, low-relief surface that dips to the west at 1° (Fig. 1B). The Chagoopa Plateau is an intermediate “bench” surface,
restricted to the major river canyons, inset into the Boreal Plateau (Webb, 1946; Jones, 1987). Associated with each surface are prominent knickpoints along major rivers. Knickpoints along the North Fork Kern River, for example (Fig. 1F), occur at elevations of 1600–2100 m and 2500–3300 m. Clark et al. (2005b) proposed that the Boreal plateau is a relict of a once regionally extensive low-elevation late Cretaceous landscape that was uplifted to its current elevation in two pulses after 32 and 3.5 Ma. In their model, late Cenozoic (<32 Ma) surface uplift of ~1.5 km initiated canyon cutting along the major rivers that has not yet propagated up into the Boreal Plateau. In an earlier paper, Wahrhaftig (1965) proposed that the stepped topography of the southern Sierra Nevada was the result of a slope-weathering feedback in which gently sloping terrain stores regolith and...
enhances weathering in a positive feedback. Wahrhaftig’s model does not appear to explain the presence of the prominent river knickpoints in the region, however.

Cosmogenic nuclide measurements provide important constraints on rates of landscape evolution. Small et al. (1997) and Stock et al. (2005) measured erosion rates of ~0.01 mm/yr on bedrock-exposed hillslopes of the Boreal Plateau. Riebe et al. (2000, 2001) calculated basin-averaged erosion rates of ~0.02 mm/yr (largely independent of hillslope gradient) on the Boreal Plateau, increasing to ~0.2 mm/yr in steep watersheds draining directly into mainstem river canyons. Stock et al. (2004, 2005) also measured mainstem incision rates along the South Fork Kings and nearby rivers as they incised into the Chagoopa Plateau. These authors found incision rates to be ~0.07 mm/yr between 5 and 3 Ma, accelerating to 0.3 mm/yr between 3 and 1.5 Ma, and decreasing to ~0.02 mm/yr thereafter.

3. Numerical modeling

Modeling erosional response to uplift requires mathematical models for hillslope and bedrock channel erosion. The classic method for quantifying bedrock channel erosion, the stream-power model, assumes that bedrock channel erosion is proportional to a power of upstream contributing area (a proxy for discharge) and channel-bed slope:

\[
\frac{\partial h}{\partial t} = U - K_w A^n \left| \frac{\partial h}{\partial x} \right|^n,
\]

where \( h \) is local elevation, \( t \) is time, \( U \) is uplift rate, \( K_w \) is the coefficient of bedrock erodibility, \( A \) is drainage area, \( x \) is the along-channel distance, and \( m \) and \( n \) are constants (Howard and Kerby, 1983; Whipple and Tucker, 1999). This model is successful at predicting channel geometries in steady-state mountain belts, where slope and drainage area scale in a systematic way. Studies have found 0.5 to be a typical value for \( m/n \), which is a measure of channel concavity in steady state (Whipple and Tucker, 1999; Snyder et al., 2000).

Sklar and Dietrich (2001, 2004) proposed an alternative “saltation–abrasion” model in which bedrock channel erosion is controlled by the sediment flux delivered from upstream. Saltation abrasion is just one of several erosional processes that can occur in bedrock rivers, but it likely to be the dominant process in rivers incising into massive bedrock lithologies such as granite (Whipple et al., 2000). Sklar and Dietrich (2001, 2004) developed complex models to quantify this process, but insights into the saltation–abrasion model can be obtained by replacing drainage area with volumetric sediment flux \( Q_s \) in the stream-power model to obtain a sediment-flux-driven erosion model:

\[
\frac{\partial h}{\partial t} = U - K_s Q_s^{m} \left| \frac{\partial h}{\partial x} \right|^n,
\]

and introducing a new coefficient of erodibility \( K_s \). This version of the sediment-flux-driven model incorporates the “tools effect” of saltating bedload but not the “cover effect” that occurs when channels are sufficiently overwhelmed with sediment to cause a reduction in erosion rates. Eq. (2) also does not explicitly include the effects of grain size on abradional efficiency-important aspects of the saltation–abrasion model proposed by Sklar and Dietrich but not critical to this paper.

The predictions of the stream power and sediment-flux-driven models are identical for cases of uniformly uplifted, steady-state mountain belts. In such cases erosion is spatially uniform (everywhere balancing uplift) and therefore sediment flux \( Q_s \) is proportional to drainage area \( A \). The predictions of the two models are very different, however, following the uplift of an initially low-relief plateau. In such cases the stream-power model predicts a relatively rapid response to uplift because large rivers with substantial stream power drain the plateau. In the sediment-flux-driven model, however, low erosion rates on the plateau limit the supply of “cutting tools” to channels draining the plateau edge, leading to much slower rates of knickpoint retreat (Gasparini et al., 2006).

Numerical experiments illustrate the difference in behavior between the two numerical landform evolution models incorporating sediment-flux-driven erosion (Fig. 2A–F) and stream-power erosion (Fig. 2G–L) with \( m=0.5 \) and \( n=1 \) for a vertically uplifted, low-relief plateau 200 km in width. Uplift occurs at a constant rate \( U=1 \) m/k.yr for the first 1 Myr of both simulations. The values \( K_w=3 \times 10^{-4} \) kyr\(^{-1} \) and \( K_s=3 \times 10^{-4} \) (m kyr\(^{-1} \)^\(^{-1} \) used in these runs would predict identical steady state conditions if uplift was assumed to continue indefinitely. In the model, hillslope erosion occurs at a constant value of 0.01 mm/yr. Isostatic rebound was estimated by assuming regional compensation over a prescribed length scale (i.e. the flexural wavelength), assumed to be equal to the model domain in this example, and averaging the erosion rate over that length scale. The average erosion rate \( E_{av} \) was computed during each time step and isostatic rock uplift \( U_i \) rate was calculated using

\[
U_i = \frac{\rho_e}{\rho_m} E_{av}
\]
where $\rho_c$ and $\rho_m$ are the densities of the crust and mantle, respectively.

Consider first the behavior of the stream-power model (Fig. 2G–L). Uplift initiates a wave of bedrock incision in which channel knickpoints propagate rapidly into the headwaters of each drainage basin. In this model, knickpoints reach the drainage headwaters after 25 Myr and the maximum elevation at that time is nearly 3 km (Fig. 2L). Following 25 Myr, the range slowly erodes to base level. In the sediment-flux-driven model (Fig. 2A–F), knickpoint migration occurs at far slower rates due to the lack of cutting tools supplied from the slowly eroding upland plateau. In this model, knickpoints require 60 Myr to reach the drainage headwaters and the maximum elevation at that time is nearly 4 km. Incision of lowland channels drives regional rock uplift in both models. In the sediment-flux-driven model, however, the low rates of channel incision on the upland...
plateau causes surface uplift of plateau remnants to higher elevations than in the stream-power model. The magnitude of this flexural-isostatic rebound effect is equal to the inverse of the relative density contrast between the crust and mantle: $\frac{\rho_c}{(\rho_m - \rho_w)}$, which is between 4 and 5 for typical crust and mantle densities (Turcotte and Schubert, 2002). Typical flexural wavelengths in mountain belts are 100–200 km. Therefore, if the extents of “relict surface” remnants are equal to or smaller than this wavelength, canyon cutting nearby will cause substantial local surface uplift of these remnants up to 4 times higher than that of the original, regionally extensive plateau, even as the mean elevation of the range is decreasing. In the limit where the uplift-rate dependence on $K_w$ is that sediment flux, not drainage area, is the key controlling variable on erosion rates in bedrock channels. Second, the rapid knickpoint retreat in the stream-power model is difficult to reconcile with the prevalence of low-relief upland plateaus in mountain belts that have not been subject to active uplift for hundreds of millions of years (e.g. Appalachians (Ashley, 1935), Australia (Twidale, 1976)). The sediment-flux-driven model, in contrast, suggests a common geomorphic mechanism for such persistent, plateau-dominated landscapes.

4. Application to the southern Sierra Nevada

I assume an initially low-relief, low-elevation landscape similar to the Boreal Plateau of the modern Sierra Nevada. However, no explicit assumption is made about the timing of initial uplift of the range. The initial topography input to the model respects the modern drainage architecture of the southern Sierra Nevada but replaces the modern stepped topography with a uniformly low-relief, low-elevation landscape (Fig. 3A). To construct the initial topography, a constant base-level elevation of 200 m was first identified based on the modern range-front elevation. The topography of the range above elevations of 200 m was extracted and the relief of that topography was uniformly lowered by a factor of five. The initial model topography produced in this way has a mean elevation of approximately 0.5 km, a maximum elevation of 1 km, and local relief similar to that of the modern Boreal Plateau. All pixels draining to areas north, east, and south were “masked” from the model domain so that only the evolution of west-draining rivers was considered.

Hillslopes and channels in the model are distinguished on the basis of a stream-power threshold (Tucker and Bras, 1998; Montgomery and Dietrich, 1999). If the product of slope and the square root of contributing area is greater than a prescribed threshold given by $1/X$, where $X$ is the drainage density, the pixel is defined to be a channel, otherwise it is a hillslope. The value of $X$ was chosen to be 0.005 km$^{-1}$ on the basis of observed drainage densities on the upland Boreal plateau measured with a 30-m resolution DEM. Hillslope erosion in the model occurs at a prescribed rate $E_h$ independent of hillslope gradient. Diffusion or nonlinear-diffusion models are commonly used to model hillslope evolution in numerical models, but cosmogenic erosion rates measured on the Boreal Plateau (Riebe et al., 2000) and the slope-independence of hillslope erosion in thin-regolith granitic landscapes (Granger et al., 2001) generally
supports the use of a slope-independent hillslope erosion model for hillslopes of the Boreal Plateau. The value of $E_h$ is constrained by cosmogenic erosion rates on bare bedrock surfaces of the Boreal Plateau to be approximately 0.01 mm/yr (Small et al., 1997; Stock et al., 2004, 2005). This value applies only to the upland low-relief surface, but in practice hillslope pixels in the model only occur on the upland landscape because of the large gradients and/or drainage areas that develop in the incised river canyons below the upland plateau. Hillslopes play an important role in the sediment-flux-driven model by supplying sediment to propagate knickpoints, but the model results are not sensitive to the particular value of $X$ as long as it is within a reasonable range of 0.002 km$^{-1}$ to 0.01 km$^{-1}$.

Bedrock channel erosion occurs in all channel pixels during each time step of the model according to Eqs. (1) or (2). Range-wide surface uplift in the model was assumed to occur as uniformly with a rate $U = 1$ m/kyr during each uplift pulse. Slope-area relationships in steady-state bedrock rivers indicate that the ratio $m$/$n = 0.5$ is typical (Whipple and Tucker, 1999; Snyder et al., 2000). This $m/n$ ratio implicitly includes the effect of increasing channel width with increasing drainage area. In these model experiments, I used $n$ values of 1 and 2, scaling $m$ accordingly to maintain a constant $m/n = 0.5$ ratio. Sediment flux is computed in the sediment-flux-driven model using erosion rates computed during the previous time step. Sediment flux and drainage area are both routed downstream on hillslopes and in channels using a bifurcation-routing algorithm that partitions incoming fluxes of sediment or drainage area between each of the downstream neighboring pixels, weighted by slope. Time steps in the model are variable (i.e. small time steps are needed when stream gradients are high following uplift in order to ensure numerical stability) but range from about 1 to 10 kyr.

Geomorphic and stratigraphic observations in the northern Sierra Nevada suggest that isostatic rock uplift takes place by westward tilting along a hinge line located near the range front. Rock uplift in the southern Sierra Nevada is accommodated by a series of normal
faults. Despite this difference in structural style, the large-scale pattern of erosional unloading east of the mountain front and depositional loading west of the mountain front suggests that the large-scale pattern of flexural-isostatic uplift in both the northern and southern Sierra occurs by westward tilting (Chase and Wallace, 1988; Small and Anderson, 1995). Therefore, I assumed that isostatic uplift in the model application to the southern Sierra Nevada has a mean value equal to Eq. (3) but is spatially distributed using a linear function of distance from the range crest:

$$U_i = \frac{L}{\langle x_c \rangle} \left( 1 - \frac{x_c}{L} \right) \frac{\rho_c}{\rho_m} E_{av}$$

where $x_c$ is the distance from the range crest along an east-west transect, $L$ is the distance between the crest and the hinge line, and $\langle x_c \rangle$ is the average distance from the range crest. I assumed $L=100 \text{ km}$ (Unruh, 1991) and $\rho_c/\rho_m=0.8$.

Calibration of $K_w$ and $K_s$ was performed for each uplift history by forward modeling the response to a late Cenozoic uplift pulse and adjusting the values of $K_w$ and $K_s$ to match the maximum incision rate of 0.3 mm/yr measured along the South Fork Kings River (Stock et al., 2004, 2005). For each uplift history considered, the values of $K_s$ and $K_w$ were varied by trial and error until the maximum value of late Cenozoic incision matched the measured value of 0.3 mm/yr on the South Fork Kings River to within 5%. Late Cenozoic incision rates in the model increase monotonically with $K_w$ and $K_{\text{in}}$, so trial values of $K_s$ and $K_w$ were raised (or lowered) if predicted incision rates were too low (or high). All uplift histories I considered assumed two pulses of range-wide surface uplift with the late Cenozoic phase constrained to occur on or after 10 Ma and on or before 3 Ma to be consistent with geomorphic and cosmogenic constraints. The magnitude of uplift during each pulse was tentatively constrained using the offset of the two major river knickpoints of the North Fork Kern River (Fig. 1F), which suggests an initial pulse of $\sim 1 \text{ km}$ surface uplift corresponding to the upstream knickpoint along the Kern and other mainstem rivers, followed by a $\sim 0.5\text{-km}$ pulse corresponding to the downstream knickpoint. Forward modeling indicates that knickpoint relief in the model remains nearly equal to the magnitude of the range-wide surface uplift that created the knickpoint, except for late stages of the model when the uplift of isolated plateau remnants can greatly outpace stream incision. Best-fit uplift histories were determined by trial and error, visually comparing the model-predicted topography with those of the modern Sierra Nevada. The spatial extents of the Chagoopa and Boreal Plateaux and the locations of the major river knickpoints were the landscape features that were most sensitive to surface uplift history. Uplift pulses that begin too early wipe out the plateau topography so that no plateau remnants or knickpoints remain. Those that begin too late predict plateaus that are more extensive than those preserved today as well as knickpoint locations and range-crest elevations that are inconsistent with the modern southern Sierra Nevada.

In the best-fit sediment-flux-driven model experiment (Fig. 3B–D), with $n=1$ and $K_s=2.4 \times 10^{-4}$ (m kyr)$^{-1/2}$, a 1-km pulse of latest Cretaceous surface uplift was imposed from time $t=0$ to 1 Myr (out of a total duration of 60 Myr) and a second 0.5-km pulse of late Miocene surface uplift was imposed from $t=50$ to 50.5 Myr. Model experiments with $n=2$ were also performed and yielded broadly similar results for the overall topographic form of the range, but shapes of longitudinal profiles produced by the model with $n=2$ were a poor match to the observed knickpoint shapes (e.g. Fig. 5D). The time progression of model topography at $t=20, 40$, and 60 Myr (Fig. 3B–D) illustrates the model evolution in comparison to the modern topography (Fig. 3E). Channel incision rates at three points along the Kings and South Fork Kings Rivers as a function of time (Fig. 4A) record two pulses of incision. Location 2 is of particular significance since that location coincides with the model calibration datum along the South Fork Kings River where Stock et al. (2004, 2005) measured maximum Plio-Quaternary incision rates of 0.3 mm/yr. Incision rates in the model decrease with increasing distance upstream. Knickpoint propagation is also slower following the first phase of uplift compared to the second phase of uplift (Fig. 4A). Both of these behaviors can be associated with the “cutting-tools” effect of the sediment-flux-driven model. First, larger drainages have more cutting tools to wear down their beds, thereby increasing incision rates and knickpoint migration rates. Second, knickpoint propagation occurs slowly following the first pulse of uplift because of the lack of cutting tools supplied by the low-relief upland plateau. Canyon cutting has affected only a very small percentage of the landscape even as late as 20 Myr following the initial uplift (Fig. 3B) in this model. In contrast, knickpoint and escarpment retreat following the second phase of uplift is enhanced by the upstream sediment supply associated with the first phase of uplift. As a result, the second knickpoint travels almost as far upstream as the first knickpoint in less than 20% of the time.

The model results with sediment-flux-driven erosion (Figs. 3B–D and 4A–B) also provide several consistency checks that aid in confidence building. First, the
maximum incision rate observed from 3 to 1.5 Ma along the South Fork Kings River was preceded and followed by much lower rates of incision (∼0.02–0.05 mm/yr) from 5 to 3 Ma and 1.5 Ma to present (Stock et al., 2004, 2005). The model predicts the same order-of-magnitude decrease in incision rates before and after knickpoint passage. The hillslope erosion rate $E_h$ was prescribed to be 0.01 mm yr$^{-1}$ based on measured cosmogenic erosion rates. The model predicts a range of erosion rates on the upland surface from a minimum of 0.01 mm/yr (on hillslopes) to a maximum value of 0.03 (in channels) (Fig. 4A, inset plot). As such, basin-averaged erosion rates on the upland landscape is approximately 0.02 mm/yr in the model, consistent with those measured cosmogenically.

Mean surface elevation increases during active uplift but otherwise decreases (Fig. 4B). Maximum elevation continually increases through time to a final value of over 4 km as a result of isostatic uplift of Boreal Plateau remnants driven by canyon cutting downstream. Mean erosion and uplift rates show a gradual increase in mean erosion rate over a 40 Myr period to a maximum value of approximately 0.05 mm/yr (Fig. 4B). Isostasy replaces 80% of that erosion as rock uplift, as prescribed by Eq. (4). The model is also able to predict details of the Sierra Nevada topography (along the North Fork Kern River, for example, compare Fig. 5 with Fig. 1C–F). The model predicts the elevations and extents of the Chagoopa and Boreal plateaux as well as the approximate locations and shapes of the two major knickpoints along the North Fork Kern River. Model results with $n=2$, however, yield broad, cuspate knickpoints that are a poor match to the observed knickpoint shapes (Fig. 5D).

Uplift in the best-fit stream-power model, with $n=1$ and $K_w=8 \times 10^{-5}$ kyr$^{-1}$, occurs as two pulses: a 1-km pulse in the late Eocene (35 Ma) and a 0.5-km pulse in the late Miocene (7 Ma) (Figs. 3F–H and 4C–D). In this model, knickpoint migration occurs at a similar rate in the first and second pulses of uplift, as illustrated by the equal durations between knickpoint passage at each of the three locations following the first and second pulses (Fig. 4C). Knickpoint propagation occurs at similar rates in the two pulses because they are a function of drainage area (which is constant through time), not sediment flux.
The rates of vertical incision are lower following the second uplift pulse because the knickpoint has a gentler grade due to the smaller magnitude of uplift.

5. Discussion

The results of this paper support two possible uplift histories corresponding to the best-fit results of the sediment-flux-driven and stream-power models. Two additional constraints, however, point to the conclusions of the sediment-flux-driven model, i.e. that range-wide surface uplift occurred first from a low-elevation, low-relief landscape in the latest Cretaceous. First, field observations support the hypothesis that sediment abrasion is the dominant erosional process in granitic landscapes (Whipple et al., 2000), hence a sediment-flux-driven model is most appropriate for the Sierra Nevada. Second, the fact that $\mathbf{K_w}$ is observed to be a function of uplift rate (Snyder et al., 2000; DeLong et al., 2007), supports the greater applicability of a sediment-flux-driven model in mountainous landscapes generally.

Accumulation rates in the southern Great Valley further support the results of the sediment-flux-driven model. Accumulation rates were high (0.1–0.5 mm/yr) from 100–55 Ma, low (<0.03 mm/yr) from 55 Ma to 10 Ma, and moderately high (~0.05–0.2) again from ~10 Ma to the present (Wakabayashi and Sawyer, 2001). At first glance this may appear to imply limited erosion in the Sierra Nevada from 55–10 Ma if one assumes that accumulation rates in the basin are proportional to upstream erosion rates (Wakabayashi and Sawyer, 2001). Sedimentary basin accumulation is usually controlled by changes in topographic loading at the basin margins, however. When subsidence and accumulation rates were high in the southern Great Valley, the adjacent Sierra Nevada was most likely gaining in mean elevation, increasing its load on the crust in order to drive flexural subsidence and hence generate accommodation space. When accumulation rates were low, the mean elevation of the Sierra Nevada was likely undergoing minor change, with sediment eroded from the range bypassing the basin due to the lack of accommodation space. Accumulation rates increased again only when the mean elevation of the Sierra Nevada rose again in late Miocene time. The accumulation history in the Great Valley, therefore, is broadly consistent with the surface uplift history inferred from the sediment-flux-driven model.

Both the stream-power and sediment-flux-driven models fail to reproduce the observed topography in at least one major respect: they do not reproduce persistent hanging valleys along the lower reaches of the major

Fig. 5. Illustrations of the topography predicted by the best-fit sediment-flux-driven model results for the North Fork Kern River basin (i.e. 1-km of latest Cretaceous uplift and 0.5 km of late Cenozoic uplift), illustrating broadly similar features to those of the actual North Fork Kern River basin (Fig. 1C–E). (D) Longitudinal profile of the best-fit sediment-flux-driven model with $\mathbf{n} = 2$. In this case, the model produces broad cuspatke knickpoints that are not consistent with observed longitudinal profiles.
river canyons. Crosby et al. (2005) argued that very steep bedrock channels (i.e. bed slopes near vertical) lose abrasional efficiency because of the oblique angle of incidence between the saltating bedload and channel bed in such steep channels. This “cosine effect” promotes the formation of hanging valleys in New Zealand and Taiwan (Wobus et al., 2006). Including this effect in the model would likely improve its realism but would also require a much more finely resolved grid in order to recognize very steep channels in the model. The cosine effect would most likely lengthen the time required for knickpoints to propagate in the model (because hanging valleys supply less sediment flux to drive downstream incision), thereby pushing back slightly the timing of the uplift pulses inferred from the model.

Uncertainty in the western extent of Sierran uplift translates into uncertainty regarding the timing of initial uplift. In the model, late Cretaceous surface uplift was prescribed to begin at 60 Ma and cease at 59 Ma, but uplift could have begun significantly earlier if the uplift rate was lower than the 1 mm/yr assumed here, if uplift extended further west, and/or if past upland erosion rates (hence sediment fluxes and knickpoint migration rates) were lower than modern rates. The purpose of the model was to test end-member scenarios for late Cretaceous versus late Cenozoic uplift. Given the uncertainties, the model cannot resolve the detailed history of late Cretaceous uplift, and the 60 Ma timing I infer based on modeling should only be considered a minimum age for the cessation of late Cretaceous uplift. The interpretations of House et al. (1997, 1998, 2001) are relevant to this point. Using similar thermochronological data as Clark et al. (2005b), House et al. concluded that most of the incision of the modern mainstem rivers had been achieved by late Cretaceous time at a distance of approximately 30 km from the range front. The results of this paper do not support such ancient uplift/canyon cutting, but they could be reconciled with that interpretation if past upland erosion rates were lower than today, thereby slowing past rates of knickpoint propagation relative to modern rates.

6. Conclusions

The results of this paper are consistent with the basic geomorphic interpretation of Clark et al. (2005b) that the upland plateaux and associated river knickpoints of the southern Sierra Nevada likely record two episodes of range-wide surface uplift totaling approximately 1.5 km. The results presented here, however, suggest that the initial 1-km surface uplift pulse occurred in latest Cretaceous time, not late Cenozoic time. In the sediment-flux-driven model, the slow geomorphic response to the initial uplift phase is caused by the lack of cutting tools supplied by the slowly eroding upland Boreal Plateau. If the behavior of the sediment-flux-driven model is correct, the model results suggest that 32 Myr (i.e. the time since onset of Sierra Nevada uplift, according to Clark et al. (2005b)) does not afford enough time to propagate the upland knickpoint to its present location. The self-consistency of the model results provide confidence in this interpretation. The model correctly reproduces details of the modern topography of the range, including the elevations and extents of the Chagoopa and Boreal Plateaux and the elevations and shapes of the major river knickpoints. The timing of the upwelling events inferred from modeling is broadly consistent with late Cretaceous and late Miocene episodes of crustal delamination (Saleeby et al., 2003).

Low-relief, high-elevation “relict landscapes” are recognized in many of the world’s mountain ranges (e.g. Clark et al., 2005a,b; Barke and Lamb, 2006; Grujic et al., 2006). The past few years have seen renewed interest in using these surfaces as surface uplift indicators. The results of this paper, however, indicate that relict landscapes can persist for tens of millions of years or more, and that thermochronologic ages providing a minimum age for the onset of incision do not necessarily coincide with the onset of range-wide surface uplift.

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References


