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Notes

Numerical modeling of fluvial strath-terrace formation in response to oscillating climate

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ABSTRACT

Many river systems in western North America retain a fluvial strath-terrace record of discontinuous downcutting into bedrock through the Quaternary. Their importance lies in their use to interpret climatic events in the headwaters and to determine long-term incision rates. Terrace formation has been ascribed to changes in sediment supply and/or water discharge produced by late Quaternary climatic fluctuations. We use a one-dimensional channel-evolution model to explore whether temporal variations in sediment and water discharge can generate terrace sequences. The model includes sediment transport, vertical bedrock erosion limited by alluvial cover, and lateral valley-wall erosion. We set limits on our modeling by using data collected from the terraced Wind River basin. Two types of experiments were performed: constant-period sinusoidal input histories and variable-period inputs scaled by the marine $\delta^{18}O$ record. Our simulations indicate that strath-terrace formation requires input variability that produces a changing ratio of vertical to lateral erosion rates. Straths are cut when the channel floor is protected from erosion by sediment and are abandoned—and terraces formed—when incision can resume following sediment-cover thinning. High sediment supply promotes wide valley floors that are abandoned as sediment supply decreases. In contrast, wide valleys are promoted by low effective water discharge and are abandoned as discharge increases. Widening of the valley floors that become terraces occurs over many thousands of years. The transition from valley widening to downcutting and terrace creation occurs in response to subtle input changes affecting local divergence of sediment-transport capacity. Formation of terraces lags by several thousand years the input changes that cause their formation.

Our results suggest that use of terrace ages to set limits on the timing of a specific event must be done with the knowledge that the system can take thousands of years to respond to a perturbation. The incision rate calculated in the field from the lowest terrace in these systems will likely be higher than the rate calculated by using older terraces, because the most recent fluvial response in the field is commonly downcutting associated with declining sediment input since the Last Glacial Maximum. This apparent increase in incision rates is observed in many river systems and should not necessarily be interpreted as a response to an increase in rock-uplift rate.

Keywords: fluvial features, landscape evolution, modeling, river terraces, sediment supply.

INTRODUCTION

Fluvial strath-terrace sequences record river incision through time and the timing of environmental perturbations to fluvial systems. Fluvial strath terraces are abandoned river flood plains with thin mantles of alluvium overlying a beveled bedrock platform (Fig. 1). Conceptually, strath-terrace formation requires changes in the ratio of vertical incision to lateral planation in a downcutting stream. Gilbert (1877) postulated that as sediment load nears stream-transport capacity, vertical erosion rates decrease, making lateral planation relatively more important. Flood plains are the result of this lateral planation, and strath terraces are vestiges of these flood plains, formed as river downcutting leads to flood-plain abandonment (Gilbert, 1877). Mackin (1937) proposed that incipient terraces form during periods of river stability—i.e., graded conditions—when lateral planation of valley walls produces wide valley floors. Graded conditions occur when a river can transport all of the imposed sediment load and is achieved through a balance between sediment load and size, discharge, and channel slope (Mackin, 1937). According to Mackin (1937), the width of the valley floor is related to the duration of the graded period. Both Gilbert (1877) and Mackin (1937) suggested that the beveling of the bedrock strath and the deposition of the alluvial mantle occurred simultaneously and that the channel geometry remained stable while the channel migrated across the valley floor. The beveled strath and thin alluvial cover distinguish strath from fill terraces, which have no underlying bedrock platform and can be many tens of meters thick.

These hypotheses suggest that strath formation requires alternation between ''graded'' periods of extensive lateral planation, and ''nongraded'' periods of vertical incision. Rivers may obtain graded conditions when available stream power equals the critical power, defined as the stream power necessary to transport enough sediment to maintain grade (Bull, 1979, 1990). Under these conditions incision rates (i.e., rates of vertical erosion) are at a minimum, allowing extensive valley widening by lateral planation. Strath terraces are created when the available stream power becomes greater than this critical power, leading

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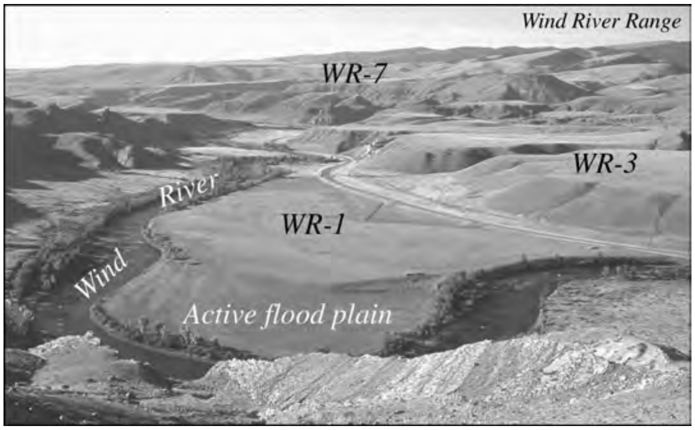

Figure 1. Photograph looking downstream along the Wind River below Dubois, Wyoming. The width of the view represents ~1 km at the center. Above the active flood plain, three terrace levels (ages: WR-1, ca. 20 ka; WR-3, ca. 120 ka; and WR-7, ca. 600 ka) mapped by Chadwick et al. (1997) can be seen along the course of the river.

to incision and flood-plain abandonment (Bull, 1979). In this conceptual model, switching between these two states can occur by altering the available stream power (e.g., changing discharge or slope), and/or the critical power (e.g., sediment load or size; Bull, 1979). Available stream power may be increased, and hence, terraces may be created by incision, through basin-scale changes such as base-level fall, hydrologic changes induced by climate change, and/or increased tectonic activity (e.g., Merritts et al., 1994; Pazzaglia et al., 1998).

The well-preserved strath-terrace sequences found in many river systems of western North America record discontinuous incision into bedrock throughout the late Quaternary (e.g., Howard, 1970; Patton et al., 1991; Reheis et al., 1991; Ritter, 1967). Their widespread occurrence implies response to a common regional forcing, and field data indicate climate oscillation between glacial and interglacial conditions, superimposed on long-term uplift and/or base-level lowering, may be responsible. Many terrace deposits indicate transport of larger grain sizes by the terrace-forming rivers compared to their modern counterparts, suggesting greater peak discharges and/or more effective headwater erosion during valley-widening periods than at present (Baker, 1974; Pierce and Scott, 1982). In addition, some terrace deposits suggest that terrace-forming rivers were braided, unlike their modern meandering counterparts (e.g., Baker, 1974; Jackson et al., 1982; Ritter, 1967; Sinnock, 1981). Braided-stream conditions suggest greater sediment supply (particularly bedload) and larger, more flashy channel flows (e.g., Bridge, 1993; Leopold and Wolman, 1957; Schumm and Khan, 1972; Smith and Smith, 1984).

On the basis of this evidence, valley widening and strath formation are proposed to occur during periods of maximum glaciation (e.g., Baker, 1974; Bull, 1991; Howard, 1970; Ritter, 1967; Sinnock, 1981), although glaciers within the headwaters of a river system are not required (e.g., Pierce and Scott, 1982). Valley widening and strath formation are initiated by increased sediment supply and hydrologic changes associated with a larger and later-melting snowpack, reduced evaporation and infiltration capacity, extensive periglacial processes, and reduced vegetative cover during cooler climate conditions (Pierce and Scott, 1982). Incision and terrace formation by abandonment of flood plains presumably occur during interglacial periods or during the initial transition from glacial to interglacial conditions (Sinnock, 1981). Church and Ryder (1972) and Jackson et al. (1982) have proposed that fluvial extraction of sediments created during glacial climates may continue for an extended time, called the paraglacial period, after the transition toward interglacial conditions has begun. Renewed incision and strath-terrace formation may therefore occur well after this transition.

The age and height of paired strath-terrace sequences provide the primary data for estimating the incision history and incision rates in fluvial systems (e.g., Pazzaglia et al., 1998). In western North America, incision histories deduced from precisely dated strath-terrace sequences are rare, but many of these sequences indicate accelerating incision rates during the Quaternary (e.g., Hancock et al., 1999; Patton et al., 1991; Reheis et al., 1991; Repka et al., 1997). The cause for this acceleration is not clear, but could indicate regional intensification of climate, base-level, and/or tectonic change. Alternatively, Mills (2000) has proposed that apparent late Cenozoic acceleration of incision rates deduced from many terrace sequences in the southeastern United States is an artifact of averaging over longer periods as terrace age increases. In our modeling, we attempt to test this hypothesis.

The timing, duration, and mechanisms of strath-terrace formation are difficult to infer solely from field observations because terrace sequences represent incomplete records, are difficult to date, and formed during fluvial conditions that differ from the present. Here we explore strath-terrace formation in response to climate oscillation by using a numerical stream-profile simulation. This approach differs significantly from previous modeling efforts focused on the generation of fluvial terraces (e.g., Boll et al., 1988; Veldkamp, 1992; Veldkamp and Vermeulen, 1988) because we rely on physically based rules and a realistic set of stream inputs. First, we discuss the model framework and rule set, the constraints on initial and boundary conditions, and our modeling strategy. Second, we discuss the results and their implications for terrace formation and interpretation. Specifically, we explore the following questions: (1) Why do terraces and terrace sequences form? (2) How is terrace formation related in time to forcing mechanisms? (3) How long does it take to form a strath terrace? (4) What past climatic episodes are most likely to leave a terrace record? and (5) What implications do the answers have for interpreting fluvial terrace sequences?

MODELING APPROACH

We use a one-dimensional finite-difference code simulating the evolution of the long valley profile of a river in the face of temporally varying inputs of water and sediment. We incorporate (1) sediment transport, (2) vertical incision of bedrock at the base of the active channel, and (3) lateral planation of the valley walls. Our incorporation of lateral planation differentiates our approach from previous stream-evolution models (see Howard et al., 1994) and allows us to investigate the conditions conducive to generation of strath terraces. Aside from lateral planation, our goals and physical rules are similar to those of Tucker and Slingerland (1997), but our focus is solely on the channel response to climatic perturbations.

Climatic forcing may be manifested in fluvial systems by changes in discharge, sediment load, and sediment size. Our model rules must therefore explicitly incorporate grain size and water discharge. We assume that transport of sediment is dictated by the local bed shear stress, and erosion of bedrock, either of the wall or of the bed, is related to the available energy per unit area of bed, or unit stream power (e.g., Whipple and Tucker, 1999). The one-dimensional model tracks sediment thickness, bedrock erosion, and valley width at evenly spaced channel cells along a profile, analogous to channel reaches. We use a staggered grid to enhance model stability. The thickness of the sediment, water discharge, channel elevation, and width of the channel and the valley floor are tracked at the center of each channel cell. The physical quantities that dictate fluxes between the cells, such as slope and shear stress, and the fluxes themselves, are calculated at the edges of the cells.

Sediment Transport

Sediment mass within each cell is conserved, as expressed by the continuity equation:

$$\frac{dh}{dt} = -\frac{1}{w}\frac{dQ_s}{dx}, \tag{1}$$

where h is sediment thickness, t is time, Q_s is volumetric sediment-transport rate, x is the downstream distance, and w is either valley width, w, or channel width, w_c. If sediment flux decreases in the downstream direction (i.e., dQ_s/dx is negative, which will cause sediment deposition), deposition is specified to occur evenly over the active valley width, w. Here the active valley width is defined as the region at nearly the same elevation as the channel, and across this width the channel migrates. Hence, channel migration deposits sediment across the active valley floor. When sediment flux increases downstream (i.e., if dQ_s/dx is positive, which will cause sediment erosion), sediment is removed only from the active channel width, w_c. Taken together, these rules allow sediment to be sequestered by deposition on wide valley floors and allow these mantled valley floors to be preserved during subsequent channel incision.

We must therefore formulate a model algorithm for sediment discharge, and we follow the dominant-discharge (i.e., geomorphically effective) strategy of Tucker and Slingerland (1997). Most sediment-transport equations suggest that the transport capacity, Q_s, depends on the local shear stress, τ,

$$\tau = \rho g H S, \tag{2}$$

where ρ is water density, g is gravitational acceleration, H is water depth, and S is energy slope, here assumed equal to channel-bed slope. Equation 2 is valid for steady, uniform flow. Both the channel slope and the water depth must be known in order to calculate the shear stress and, hence, the sediment discharge. The channel slope, S, at each cell edge is calculated by differencing the bed elevation, z, from the adjacent channel cells and dividing by the cell spacing, dx.

We calculate the flow depth, H, in a cell from the water discharge, Q. The effective discharge, Q, defined as the flow having the right combination of frequency and magnitude to control channel form, is related to the drainage area, A, through a power-law relationship (e.g., Dunne and Leopold, 1978),

$$Q = bA^c, \tag{3}$$

where b and c are empirical constants. Temporal variation in discharge is prescribed by varying b and/or c. Channel width, w_c, is calculated by using the hydraulic geometry relationship

$$w_c = dQ^e, \tag{4}$$

where d and e are again empirical constants. In our models, appropriate values for b, c, d, and e in equations 3 and 4 are taken from the Wind River, Wyoming, and are typical of those obtained in numerous river systems (Leopold and Maddock, 1953).

Flow depth in a channel is given by

$$H = \frac{Q}{vw_m}, \tag{5}$$

where v is mean flow velocity and w_m is mean flow width. Assuming a roughly rectangular channel, $w_m \approx w_c$. The mean velocity may be estimated by using the Manning equation

$$v = \frac{kH^{2/3}S^{1/2}}{n}, \tag{6}$$

where k is a constant and n is the Manning roughness coefficient. Furthermore, we have assumed a wide, shallow channel such that w_c » H, allowing substitution of flow depth H for the hydraulic radius.

By combining equations 3, 4, 5, and 6, we obtain the relationship between flow depth, H, and measurable quantities:

$$H = \left(\frac{bA^c n}{w_c}\right)^{3/5} S^{-3/10}. \tag{7}$$

Finally, by combining equations 2 and 7, bed shear stress may be predicted with the desired explicit dependence on the discharge:

$$\tau = \rho g\left(\frac{bA^c}{w_c}\right)^{3/5} n^{3/5}S^{7/10}. \tag{8}$$

To predict the volumetric sediment flux, Q_s, we use a variation of the Bagnold bedload sediment-transport equation (see Slingerland et al., 1994),

$$Q_s = \frac{Bw}{(\rho_s - \rho)\rho^{1/2}g}(\tau - \tau_c)(\tau^{1/2} - \tau_c^{1/2}), \tag{9}$$

where B is a dimensionless constant (~10), τ_c is the critical shear stress required to transport a given grain size, and ρ_s is the density of the sediment grains. The τ_c for sediment entrainment is estimated by using the Shields parameter, and this is the means by which dependence upon grain size is brought into the problem. The sediment discharge, Q_s, is determined by using the shear stress (equation 8) calculated at the edge of each channel cell. The divergence of the sediment-transport capacity, dQ_s/dx, may then be calculated at each channel cell center. Although equation 9 is formulated solely for bedload transport in alluvial or transport-limited rivers, bedload is likely the most critical component affecting stream incision and terrace generation, as the bedload must be in transit before rock erosion can take place. The simulated streams are largely covered by alluvium, with only patchy bedrock exposure.

Bedrock Erosion

Rules for predicting bedrock erosion at the reach scale in landscape-evolution models are based on either shear stress or the stream power per unit bed area, ω,

$$\omega = \frac{\rho g Q s}{w_c}, \tag{10}$$

where dz/dt is the vertical erosion rate (e.g., Whipple and Tucker, 1999). These rules are typically of the form

$$\frac{dz}{dt} = K(X - X_0)^r, \tag{11}$$

where X is either shear stress or stream power, X_0 is a threshold shear stress or stream power,

r is an exponent, and K is a parameter relating "excess" shear stress or stream power to erosion rate. This formulation neglects explicit consideration of the role of the sediment-supply rate in controlling erosion (Sklar and Dietrich, 2001). In this formulation, the value of K is related to rock properties that dictate the amount of rock that may be eroded per unit shear stress or stream power. Values for K are not well defined, but have been determined by erosion measurements in badland channels (Howard and Kerby, 1983) and by comparing past and present channel profiles in the Sierra Nevada, California (Stock and Montgomery, 1999). The method for choosing K for a particular setting is not known, however. Our approach is to select a K that produces geologically reasonable erosion rates, to assume $r = 1$ and $X_0 = 0$, and to use stream power per unit bed area. Results obtained by using the stream power and shear stress rules do not differ substantially, as shown theoretically by Whipple and Tucker (1999). We use separate rules that predict vertical rock erosion (i.e., incision) rates and horizontal rock erosion (i.e., lateral planation) rates.

Rock-Incision Rates

Potential vertical rock-erosion rates, $(dz/dt)_p$, is expressed by

$$\left(\frac{dz}{dt}\right)_p = K_v\omega, \tag{12}$$

where K_v is a "hardness" parameter for vertical rock erosion. This equation predicts only the maximum potential erosion rate because rock incision can take place only when sediment does not protect that rock from erosion. Field observations suggest that bedrock erosion can occur while there is a finite mean sediment thickness within a reach, as sediment cover is often variable (e.g., Leopold and Maddock (1953). Sediment may be transported and stored within a reach even while bedrock erosion is occurring (Shepherd and Schumm, 1974; Howard and Kerby, 1983) and may be required for this erosion to take place as it provides the abrasive tools (e.g., Hancock et al., 1998; Sklar and Dietrich, 2001; Whipple et al., 2000). We do not incorporate the possibility that there may be an optimal sediment-supply rate, above and below which erosion rates decline (Sklar and Dietrich, 2001).

To accommodate these observations, we add to equation 12 a sediment-scour rule. This is designed to acknowledge that a finite but thin sediment cover can be scoured away in high-transport conditions, exposing the underlying rock to erosion. We assume that the depth of sediment scour is related to the sediment-transport capacity, Q_s, and the channel width, w_c, and may not be uniform. We use the potential sediment-transport rate, Q_s, calculated by using equation 9, to estimate a sediment-scour depth scale, L,

$$L = \frac{sQ_s}{w_c}, \tag{13}$$

where s is a constant with dimensions of time per length, and w_c is channel width.

It is reasonable that the probability of scouring to bedrock increases as the scour depth scale, L, increases, and decreases as the sediment thickness, h, increases. Following Howard (1998), we hypothesize that the probability, F, for rock exposure and, hence, of rock incision, decreases exponentially as sediment thickness increases:

$$F = e^{-(h/L)}. \tag{14}$$

Appropriately, the probability of bedrock erosion is therefore 1 if $h = 0$ and approaches zero as $h \gg L$. To calculate the actual vertical bedrock-incision rate in a cell, dz/dt, we combine equation 12 and equation 14 to yield

$$\frac{dz}{dt} = FK_v\omega. \tag{15}$$

Although speculative, this formulation accommodates several field observations, including (1) the need to scour to the base of the sediment layer in order to erode rock, and (2) the likelihood that scour depths will vary within the channel. This formulation also recognizes that rock-erosion rates should increase as Q_s increases and decrease as the sediment thickness increases.

Lateral Planation Rates

Although the processes of erosion of river banks in self-formed, alluvial river channels have been the subject of extensive investigation (e.g., Andrews, 1982), the widening of bedrock valley walls by rivers has received much less attention. We know of only one study of lateral erosion, and this study suggests that lateral erosion rates, dw/dt, are related to $\omega^{1/2}$ (Suzuki, 1982). However, the probability of valley-wall contact was not considered, as we discuss subsequently.

To proceed, we make the assumption that the potential rate of valley-wall erosion is analogous to vertical rock erosion and is related to stream power per unit bed area,

$$\left(\frac{dw}{dt}\right)_p = K_w\omega \tag{16}$$

applied to the channel margins, where $(dw/dt)_p$ is the potential valley-widening (not channel-widening) rate and K_w is a susceptibility constant that relates stream power to wall erosion. We acknowledge that this critical component of our model cannot be justified rigorously. Investigation of the valley-widening processes in rock-floored channels and of the controls on widening rates is sorely needed. However, lateral planation that far exceeds vertical incision over time is a key field observation that must be reproduced in this model. For example, in the Wind River basin (Fig. 1), terrace widths require average lateral planation rates to be several orders of magnitude greater than vertical rates. As discussed subsequently, the model succeeds in reproducing this observation.

To widen the valley, a channel must be in contact with its valley wall. Because our model is one-dimensional, we do not track the channel position explicitly. Instead, we assume that the channel position within a valley is random if sampled over a long time period in the absence of preferred positioning imposed by tilting or other external forcing. In this case, the channel width, w_c, and the valley width, w, determine the probability of valley-wall contact, W,

$$W = \frac{w_c}{w}. \tag{17}$$

The probability increases as w_c increases and as w decreases. The rate of valley-wall erosion is then determined by combining equations 16 and 17:

$$\frac{dw}{dt} = WK_w\omega. \tag{18}$$

We note that valleys should widen rapidly at first and more slowly thereafter as the probability of channel contact with the valley wall declines.

This lateral erosion is divided between the two valley walls. In order to allow asymmetry in wall erosion, we specify a symmetry constant, α, where $0 \leq \alpha \leq 1$. Rates of valley-wall erosion are $\alpha dw/dt$ and $(1-\alpha)dw/dt$, on opposite valley walls. If $\alpha = 0.5$, there is no preferred slip direction. As illustrated in Figure 1, preferred slip across the valley is an important factor in preserving terraces in some fluvial systems. In simulations presented here, we choose $\alpha = 1$, equivalent to forcing channel migration in a constant direction and maximizing terrace preservation.

MODELING STRATEGY

Our goal is to explore how a river responds to variations in sediment and water fluxes (here referred to as inputs) driven by climate oscillation. We seek those conditions that lead to terrace formation. We discuss two experiments in which these inputs were varied: (1) constant period and amplitude variation of inputs (the ''constant-period experiments''), and (2) irregular period and amplitude variations scaled by the marine oxygen isotope record (the ''variable-period experiments''). Although we are interested in the general phenomenon of strath-terrace formation, we have selected the Wind River, Wyoming, to provide limits for the model initial and boundary conditions and the magnitudes of discharge, sediment load, and sediment size. This choice does not limit the significance of the model to this river system, but ensures that we select geologically reasonable values.

Initial and Boundary Conditions

All experiments start with a straight longitudinal channel profile with a slope $S = 0.01$ and a total length of 150 km; the spacing of the model cells is 1 km. The slope is equivalent to maximum slopes measured at the kilometer scale on the oldest terrace profiles in the Wind River system. Boundary conditions must be specified at the upper and lower ends of the channel profile. The downstream boundary is lowered at 0.15 mm per model year, the average incision rate obtained from dated terraces at the lower end of the terraced reach of the Wind River near Riverton, Wyoming (see Fig. 2 *in* Chadwick et al., 1997). All sediment reaching the lowest channel cell is allowed to escape downstream, preventing aggradation at the lower boundary. Sediment is delivered only to the cell farthest upstream. The bedrock susceptibilities to erosion, K_v and K_w, are set to 2×10^{-11} m/(W·s/m^2) [= (m·W·s)$^{-1}$] in all cells and in all simulations, implying uniform rock resistance. This value yields geologically reasonable erosion rates during the simulations.

Water Discharge and Channel Geometry

The geomorphically effective—i.e., ''dominant'' (Tucker and Slingerland, 1997)—discharge is the relevant quantity for prediction of sediment transport and channel erosion (e.g., Wolman and Miller, 1960). The relationship between flow frequency and magnitude that produces the effective discharge for rock erosion in bedrock-floored rivers is not known. We assume that the effective flows equal the 2-yr-recurrence-interval peak discharge. Values for effective discharge were determined from modern gauging records on the Wind River. This discharge increases with drainage area, suggesting values of $b = 0.29$ and $c = 0.76$ in equation 3. Channel widths at various flows have been measured at many sites along the Wind River (Leopold and Wolman, 1957; Smalley et al., 1994), yielding values of $d = 3.0$ and $e = 0.55$ in equation 4.

Because effective discharge, Q, likely varies in response to changing climate conditions, Q is varied in our model. Paleohydrologic studies in western North America suggest that peak discharges in some terrace-forming streams were much greater than at present (Baker, 1974; Pierce and Scott, 1982). In the simulations, the effective discharge is increased by up to two times the modern effective discharge, which is a reasonable estimate for sustained, frequent flow events. In experiments in which water discharge is varied, the value of b, representing the effective basin runoff in equation 3, is increased by up to two times the modern value (i.e., $0.29 < b < 0.58$). The duration of effective discharge is 10 days per model year, a choice based upon the observed duration of the peak discharge of the Wind River. Although not considered here, changes in the duration of the effective part of a river hydrograph could also be important in a fluvial system.

Sediment Load

The sediment load and grain size also change in response to climate oscillations and dictate the ability of the river to access and erode bedrock. The initial grain size, D, is taken to be equal to the mean grain size measured in the modern Wind River ($D_{50} \approx 0.5$ cm, Smalley et al., 1994). The grain size is assumed to be uniform along the model channel; transport of only this grain size is considered. The minimum sediment-load input is ~5000 m^3/yr, equal to the measured average annual bedload transported by the modern Wind River near its headwaters at Dubois (Leopold et al., 1960).

Hallet et al. (1996) have shown that extensive glacial cover can increase sediment yield by up to an order of magnitude over nonglaciated or slightly glaciated basins. On the basis of this observation, the sediment-load input to the uppermost cell is increased by up to 10 times the minimum value. In the Wind River system, mean grain sizes transported by the modern river are similar to those found within terrace deposits, whereas in other river systems, mean and maximum grain sizes are several-fold larger in preserved terrace deposits (e.g., Baker, 1974; Pierce and Scott, 1982). In the experiments we report here, the grain size is kept constant. Experiments incorporating variations in the grain size used produce results comparable to sediment-supply variation.

Numerical Experiments

For the constant-period experiments, a fixed sinusoidal input variation with a period of 100 k.y. is selected to mimic the dominant period of climate cycling through glacial to interglacial conditions since ca. 800 ka. Maximum sediment input, sediment size, and water discharge are assumed to coincide with maximum glaciation. We describe first a control experiment with steady inputs, followed by three experiments: variable sediment load (variation by a factor of 10); variable discharge (variation by a factor of 2); and combined discharge and sediment-load variation. All experiments are run for 800 0000 model years. We discuss only the last 400 000 yr, by which time the channel response to the imposed initial conditions is complete.

In the variable-period experiments, the benthic marine oxygen isotope record of Imbrie et al. (1984) is used to scale the inputs. Because this record is a proxy for the extent of continental glaciation, we assume that (1) the magnitude of climate change (e.g., glacier extent) in our modeled river system parallels the continental ice volume, and (2) the relationship between climate change and sediment supply and water discharge is linear. Although these assumptions are certainly not entirely valid, we are interested primarily in exploring the river response to complex input variation. We consider two variable-period experiments: (1) variation of sediment supply by a factor of 10 and (2) variation of water discharge by a factor of 2. In all cases, the peak in the input is set to occur when the oxygen isotope curve indicates glacial conditions. In order to remove any memory of the initial conditions, the ca. 800 ka record of Imbrie et al. (1984) is repeated once, producing experiments with a total duration of 1600×10^3 model years.

RESULTS

Constant-Period Experiments

In the control experiment, the channel profile evolves toward a slightly concave-up

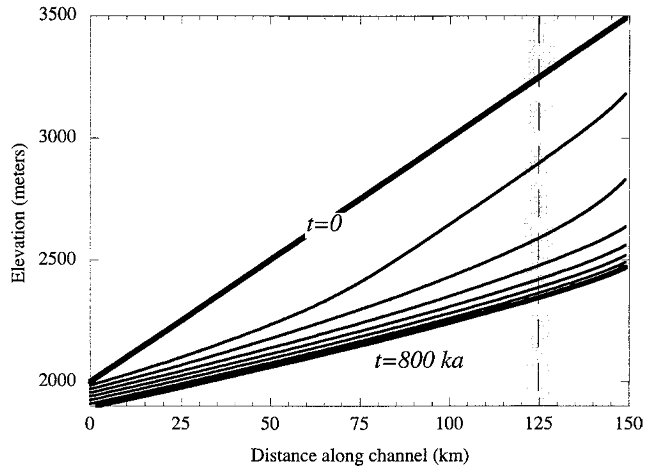

Figure 2. Successive channel profiles for every 100 k.y. model years in the control experiment, from initial conditions ($t = 0$) to simulation end ($t = 800$ ka). As in all of the constant-period experiments, the channel profile achieves a nearly uniform incision rate over the last several 100 k.y. The downstream end of the modeled profile is at 0 km.

shape. The slope in the lower half of the model space declines monotonically through time, whereas that near the top of the profile steepens slightly (Fig. 2). Both vertical and lateral erosion rates at any point within the model stream decline monotonically during the simulation. By the end of the simulation, the profile has achieved a nearly uniform erosion rate. No terraces are produced in this simulation. All other constant-period simulations produced terrace sequences, as illustrated for the sediment-load experiment in Figure 3. The six terrace levels in Figure 3 correspond to the final six input cycles completed. We discuss the controls on terrace formation, timing, and appearance, by describing the response of one representative channel cell 125 km above the downstream end of the model profile.

Vertical bedrock-incision rates, dz/dt, and lateral planation rates, dw/dt, vary through model time in all simulations in which inputs are varied (Fig. 4). The most rapid incision occurs when stream power and/or the likelihood of bedrock access is high. As a consequence, the vertical rock-incision rate is inversely proportional to variations in sediment supply (Fig. 4A), because bedrock access and the energy expended to transport sediment are

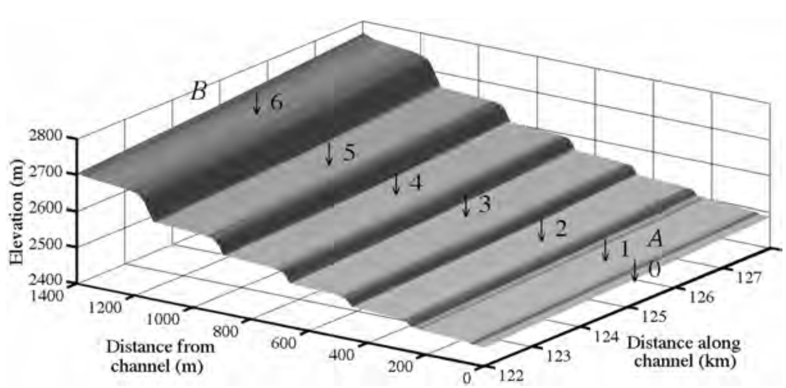

Figure 3. View of one side of the terraced-valley topography generated during the tenfold sediment-supply-variation experiment. The view extends upstream from 122 to 128 km along the model profile. The view is looking upstream and illuminated from the left. The valley floor (i.e., the river) is on the right at 0 m distance across the valley (bottom axis). In this experiment, the channel was forced to slip across the valley in one direction only; hence, the missing valley side would be a vertical wall. Each numbered terrace level was produced during one sediment-supply input cycle. All are strath terraces, with alluvium thickness of less than or approximately equal to 1 m. The topographic profile extending from A to B is shown in Figure 7.

partially controlled by sediment supply. In contrast, vertical rock-incision rates are directly proportional to changes in discharge (Fig. 4B), because increasing discharge increases available stream power.

The ratio of vertical incision to lateral planation rates, $dz/dt{:}dw/dt$, is also time dependent and is driven largely by variation in vertical incision rates (Fig. 5). Lateral planation rates are controlled almost entirely by the available stream power and are far less sensitive to variations in sediment supply than are vertical incision rates (Fig. 5). The valley width is controlled by the magnitude of this ratio and by the duration of lateral planation. Hence, wide valley floors are created during long periods when the ratio is low, when $dw/dt \gg dz/dt$. Because this ratio varies, valley width is also time dependent (Fig. 5). As envisioned by Gilbert (1877) and Mackin (1937), variation in the relative importance of lateral planation rates and vertical incision rates, here expressed as the ratio $dz/dt{:}dw/dt$, is therefore responsible for producing terrace sequences. Valley floors are widened when the ratio $dz/dt{:}dw/dt$ is low for extended periods, and terraces form when these valley floors are abandoned during subsequent incision when the ratio $dz/dt{:}dw/dt$ is high (Fig. 5, arrows show times of valley abandonment and terrace formation).

Lateral planation and therefore terrace cutting occurs when the sediment input forces $dz/dt{:}dw/dt$ to be low. In the sediment-load experiments, vertical incision rates are lowest when the input of sediment is high, allowing extensive lateral planation (Fig. 5A). This relationship corresponds to times when the sediment cover is sufficient to reduce the likelihood of scour through this cover, preventing vertical rock erosion. Declining sediment load leads to thinning of the sediment cover, allowing more rapid incision and abandonment of wide valley floors to form strath terraces. On the other hand, valleys widen when effective water discharge is low and are abandoned as increasing discharge results in more rapid vertical incision (Fig. 5B). Although the lateral planation periods necessary to form terraces may be related in time to maximum sediment load or minimum water discharge, the actual timing of terrace formation (arrows in Fig. 5) is not synchronous with these periods. Instead, in all simulations, the timing of actual terrace formation (valley-floor abandonment) typically lags these variations in the forcing by up to several tens of thousands of years.

Because sediment supply and water discharge modulate vertical incision rates differently, these inputs when combined compete to

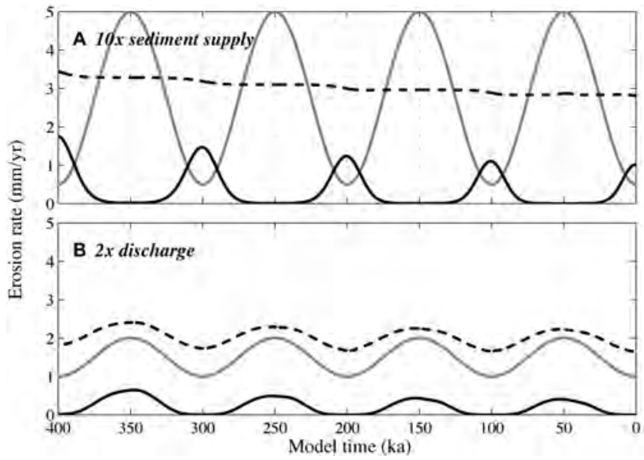

Figure 4. Vertical (solid line) and lateral (dashed line) erosion-rate histories at km 125 during constant-period variation of fluvial inputs (gray line). Simulations are (A) tenfold sediment-supply variation, and (B) twofold dominant-discharge variation. Input variation produces variable erosion rates and more significantly modulated vertical erosion rates than horizontal erosion rates. High vertical erosion rates are associated with periods most conducive to enhancing rock accessibility (i.e., low sediment supply) and/or to producing high stream power (i.e., high dominant discharge). The cycle in the ratio of vertical to lateral erosion rates is responsible for creating terraces.

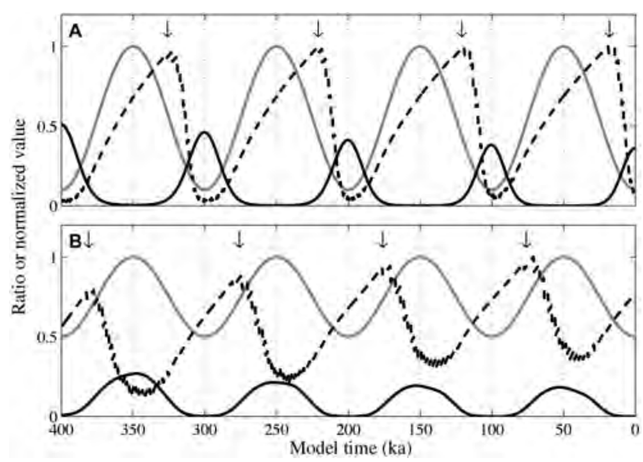

Figure 5. History of normalized sediment inputs (gray line), normalized valley-floor width (i.e., flood-plain width) at river level (dashed line), and the ratio of vertical to lateral erosion rates (solid black line) during two simulations: (A) tenfold sediment-supply variation and (B) twofold dominant-discharge variation. Inputs and valley-floor widths are normalized by dividing the value at each time by the maximum value during each simulation. The valley floor (i.e., the "flood plain") widens most significantly when vertical erosion rates are low, producing a ratio of vertical to lateral erosion that is small or zero. Flood plains are abandoned to form terraces when vertical erosion rates and the ratio of vertical to lateral erosion rates increase, leading to renewed downcutting and narrowing of the active valley. Terraces are formed at the transition from valley-floor widening to narrowing (arrows). Terrace formation does not occur at the times of either maximum inputs or minimum inputs, but instead significantly lags the timing of maximum (e.g., sediment-supply maxima) or the minimum (e.g., water-discharge minimum) inputs. In the simulations, all terraces generated are strath terraces.

control the timing of valley widening and terrace formation (Fig. 6). For the chosen amplitudes in this simulation, the sediment supply dominates over discharge in modulating the timing and magnitude of incision rates, as dz/dt is inversely proportional to the sediment supply. We illustrate with one simulation where sediment supply varies by tenfold while effective discharge varies twofold. Extensive valley widening occurs when sediment load is highest, and terraces are formed as sediment load decreases (Fig. 6). Valley widening is promoted by high effective discharge, when the lateral planation rate reaches a maximum, while vertical incision rates are low because sediment load is high and prevents access to the bedrock (Fig. 6). For this particular choice of geologically defensible input amplitudes, the variation of sediment load dictates the timing of valley widening.

The final width of a terrace is dictated by the product of the lateral planation rate with a channel-residence time at a particular elevation. We define the channel-residence time to be the time spent by the channel between successive elevation contours, here taken to be 1 m. The residence time in lateral planation periods over one simulation (that with 10× sediment-supply variation) varies from ~15 k.y. to ~60 k.y., i.e., ~15% to ~60% of each 100 k.y. cycle (Fig. 7). We note that despite this variation, the terraces produced during these periods are roughly the same width. As downcutting progresses, an increasing duration of lateral planation (low dz/dt:dw/dt) is offset by a decrease in the instantaneous lateral planation rates caused by the decline in channel slope (Fig. 7). The likelihood of wall contact (equation 17) decreases as terraces widen and also acts to limit terrace width.

Variable-Period Experiments

In the variable-period experiments, all simulations once again produce terrace sequences (Fig. 8), and the relationship between inputs and the timing and mechanisms of terrace formation is similar to that relationship for the constant-period simulations (Fig. 9). However, the irregular input variation produces an irregular channel-incision history in which the stream is unable to incise for much of each simulation (Fig. 9). Periods of lateral planation (low dz/dt:dw/dt) account for ~50%–90% of the total simulation time, implying that the typical mode of operation for the model stream system involves extended periods of lateral planation and long residence time on the flood plain interrupted by short periods of rapid incision (Fig. 10).

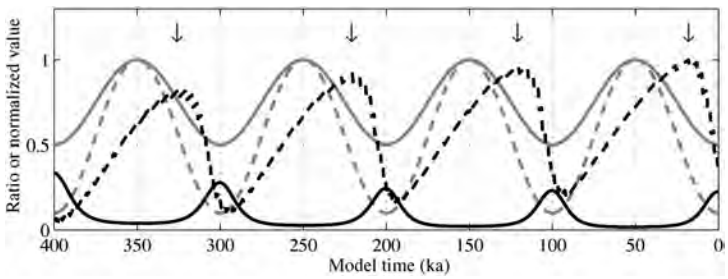

Figure 6. Ratio of vertical to lateral erosion rates (solid black line) and normalized active valley width (dashed black line) at km 125 during constant-period variation of both sediment supply (dashed gray line) and effective discharge (gray line). As shown in Figure 5, the timing of maximum vertical incision produced by these inputs is not in phase. Therefore, these inputs compete to set the timing of maximum vertical incision and, hence, terrace formation. The arrows mark the timing of valley abandonment and terrace formation. Although there is a continuum of possible combinations of these two inputs, with the field-defined amplitudes chosen for this simulation, the sediment-supply fluctuations dominate.

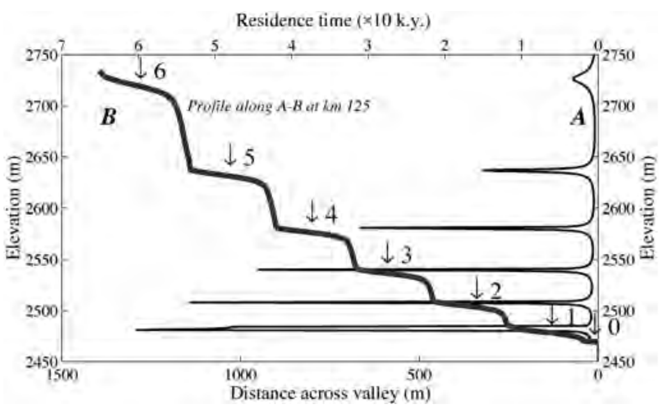

Figure 7. The topographic profile along A–B of the terraced valley illustrated in Figure 3 (thick line) and channel-residence time (see definition in text) as a function of elevation during downcutting in this simulation (thin line). Terraces are related to periods of lateral planation during long channel residence within a narrow elevation range. Residence time at individual terrace levels reaches up to many tens of thousands of years, indicating that the river spent most of the simulation forming terraces.

Unlike the constant-period experiments, not all significant input fluctuations form terraces. The terrace sequences preserved include four to six well-defined terrace levels as well as several less distinct levels (Figs. 8 and 10). The widest terraces form in response to the longest periods of high sediment load or low water discharge (Fig. 9). However, neither terrace width nor preservation perfectly records the duration and occurrence of climate events. A distinct terrace is not formed by the short-lived (~20 k.y.) fluctuation analogous to the marine isotope stage 4 event (ca. 75 ka, Fig. 9). Instead, evidence for this fluctuation is lost within the last terrace formed, created at the end of a lateral planation period lasting ~80 k.y. and spanning from isotope stage 5 to the end of isotope stage 2 (Fig. 9). Within the same simulation, three fluctuations between ca. 200 and 250 ka that are comparable to isotope stage 4 do leave three small but distinct terraces, one of which forms terrace level 3 in Figure 10.

The number of terraces preserved in a simulation depends upon the extent to which the channel migrates in a preferred direction or slips sideways in its valley. In the simulations we have presented, we forced a constant migration direction, and seven to eight terrace levels are preserved. In the absence of this imposed unidirectional migration, many terraces are cannibalized during later lateral planation periods. With no migration, at most one or two terrace levels are preserved; for example, in the sediment-variation experiments, the surviving terraces have model ages of ca. 20 and ca. 420 ka. In the absence of a preferred migration direction, terrace preservation requires that successively younger terrace levels be narrower, reflecting a shorter duration of lateral planation and/or slower lateral planation rates. In this sense, the preservation of terraces within sequences is analogous to glacial moraine survival (Gibbons et al., 1984); if a river cannot migrate primarily in a single direction, the terrace sequence will include only those terraces that are able to survive later lateral planation events. Situations that promote unidirectional channel migration include repeated river diversion (e.g., Chadwick et al., 1997) and tectonic tilting (e.g., Reheis, 1985).

Although base-level lowering in these simulations is constant, the average vertical channel-incision rate between successive terrace levels varies widely (Fig. 11). In the sediment-supply-variation experiment, rates of channel incision generally increase toward the end of the simulation; most dramatically, the average incision rate calculated from the age and elevation of the last terrace created is several times higher than any other calculated incision rate (though this increased rate seems to be more apparent than real; see discussion and Fig. 11). We reiterate that these results are from the last 800 k.y. of the total 1600 k.y. model run, and the response to initial conditions was complete before this period.

DISCUSSION

In our simulations, the conditions associated with terrace formation support many of the existing conceptual models for terrace development. A prerequisite for terrace formation is variability in the ratio of vertical incision rates to lateral planation rates, as suggested by Gilbert (1877) and Mackin (1937), among others. Wide valleys form during periods of lateral planation when bedrock exposure is limited and/or available stream power is low (i.e., the stream power nearly equals the critical power), resulting in low vertical channel-incision rates. These conditions are promoted by relatively high sediment load and/or low effective discharge. Vertical incision promotes abandonment of these valleys, producing terraces, when these factors reverse. As suggested in the introduction, sediment load and grain size are thought to have increased in many western North American streams under colder climates. Our model results therefore support the general hypothesis that valleys are widened during glacial times, when sediment production in river headwaters is generally high,

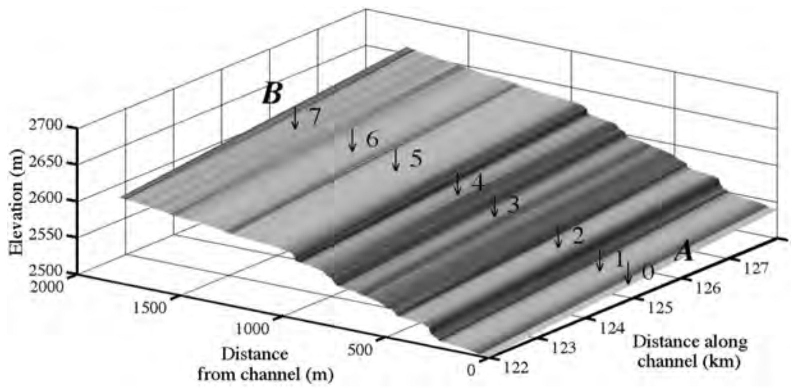

Figure 8. View of one side of the terraced-valley topography generated during the tenfold sediment-supply-variation experiment scaled by the oxygen isotope record. The view is upstream, extends from 122 to 128 km along the model profile, and is illuminated from the left. The valley floor (i.e., the river) is on the right at 0 m distance across the valley (bottom axis). In this experiment, the channel was forced to slip across the valley in one direction only; hence, the missing valley side would be a vertical wall. All terraces are strath terraces, with alluvium thickness of less than or approximately equal to 1 m. In contrast to sinusoidal variation (Fig. 3), the irregular variation of sediment supply produces terraces of variable width. The most extensive terraces (1, 2, 5, 6, and 7) are produced by lateral planation during prolonged channel residence at each terrace elevation (see Fig. 10). The topographic profile extending from A to B is shown in Figure 10.

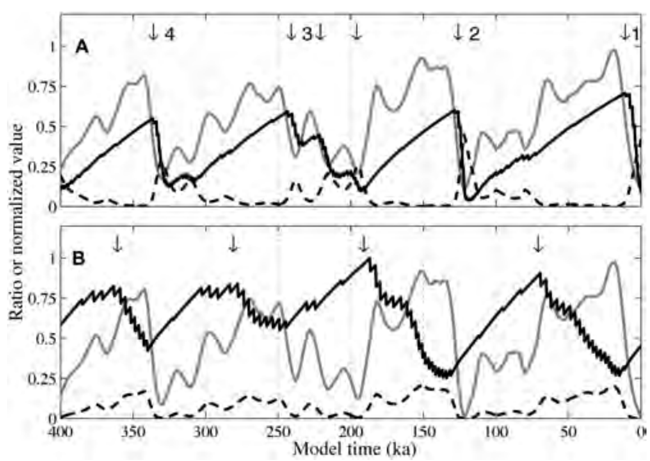

Figure 9. History of normalized inputs (gray line), normalized valley-floor width (i.e., flood-plain width) at river level (solid line), and ratio of vertical to lateral erosion rates (dashed black line) during the tenfold sediment-supply-variation experiment. Inputs and valley-floor widths are normalized as in Figure 5. Arrows indicate time of terrace formation, and the mechanism for terrace formation was as described in Figure 5. The irregular variation of inputs produces a more complex terrace sequence (Fig. 8), and terraces are not formed from all significant input variations. During the general increase in sediment supply between full interglacial (i.e., lowest sediment supply) and full glacial conditions (i.e., peak sediment supply), minor fluctuations superimposed on the larger trend do not produce a terrace. For instance, in the growth of valley width between model time ca. 120 ka and ca. 20 ka, an increase and subsequent decrease in sediment supply at model time ca. 75 ka (isotope stage 4) produces no terrace in spite of a relatively large amplitude sediment-supply change. In western North America, preservation of isotope stage 4 terraces is rare, despite a readvance of alpine glaciers at this time.

and terraces are formed by renewed incision into these valley floors during the transition to interglacial conditions.

Regardless of which input fluctuates, terrace formation lags significantly behind the timing of forcing. This lag arises because the fluvial system takes a finite period of time to pass the sediment accumulated during periods of low water discharge and high sediment input. Once the sediment cover thins at a site, bedrock becomes more accessible to erosion, allowing the incision rate there to increase. This model observation is consistent with the idea of a paraglacial recovery period (e.g., Church and Ryder (1972), during which channels adjust to the sediment loads and sizes supplied by glacial advances. If relevant to the field, this result implies that ages obtained from terraces may significantly lag the timing of maximum glaciation or cooling that drove the formation of the terrace.

If our numerical results are relevant for real river systems, the observation of long lateral planation periods interrupted by brief and rapid incision in the model suggests that interpretation of terrace age is more complex than previously recognized. If terrace surfaces are constructed during tens of thousands of years (Fig. 10), it is likely that the deposition of alluvium and cutting of the strath surface are diachronous in the cross-valley direction on an otherwise continuous terrace surface. An absolute age obtained by dating material incorporated in a terrace surface could reflect any time between the initiation of strath cutting and the timing of strath abandonment. Any numerical age should therefore be considered a maximum estimate for the timing of terrace formation, unless geologic evidence suggests otherwise. Multiple samples from the same terrace surface would help set limits on the residence time on the strath surface and would place bounds on the extent of diachroneity within the terrace material.

In field settings, the heights of dated strath terraces above modern river level are used frequently to estimate mean incision rates (e.g., Burbank et al., 1996; Chadwick et al., 1997; Repka et al., 1997). Our simulations suggest that channels spend a large fraction of time in lateral planation, and mean incision rates obtained by using terrace ages are many times lower than the maximum instantaneous incision rates within the period of averaging. In the sediment-supply simulation, for instance, calculated mean incision rates are 1.5–7 times lower than the maximum instantaneous incision rates occurring during river incision (Fig. 9). Because average incision rates may incorporate periods of little to no vertical incision

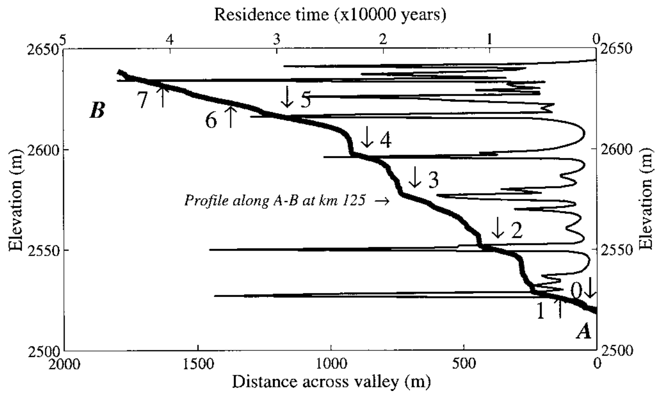

Figure 10. The topographic profile along A–B of the terraced valley illustrated in Figure 8 (thick line) and channel-residence time (see definition in text) as a function of elevation during downcutting in this simulation (thin line). Terraces are related to periods of lateral planation during long channel residence within a narrow elevation range. As in the constant-period experiments, channel-residence time at individual terrace levels reaches up to many tens of thousands of years, indicating that the river spent most of the simulation time forming terraces. The widest terraces are produced by the longest residence times, and incision between terrace levels occurs over relatively short time intervals.

of unknown duration, comparison of mean incision rates obtained in two different rivers during the same time period—or within the same river system over differing time periods—neglects the actual duration of incision. Drawing conclusions about the controls on rock-incision rates by comparing downcutting rates obtained from dated terraces should be done cautiously in light of our findings. In addition, variation in rock-incision rate is an inevitable consequence of climatically induced variation in the sediment and water inputs to the fluvial system. Changes in the mean incision rates through time as deduced from a series of terrace ages therefore do not necessarily imply long-term changes in uplift rates, but can instead simply result from variations in the sequencing and duration of input events. A stream may incise rapidly not only because it has high stream power and/or is flowing across weak rock, but also because the channel is able to transport the supplied sediment to provide sufficient access to the bed.

Many dated terrace sequences in western North America and elsewhere indicate acceleration of incision rates toward the present (e.g., Chadwick et al., 1997; Patton et al., 1991; Reheis et al., 1991). Our simulations suggest that this apparent acceleration may simply reflect the particular time at which these systems are sampled. To illustrate, in our experiments where sediment supply most strongly controls vertical erosion rates, as we suspect is the real case, the model stream is incising actively at the end of the simulation period (e.g., Figs. 4 and 9). As a consequence, the average channel-incision rate calculated from the height and abandonment age of the youngest terrace is much higher than incision rates calculated between older terrace levels (compare incision rates for terraces, Fig. 11). The increase in the average incision rate arises solely because the simulation ends when sediment input is low and when incision rates are at their highest, rather than from changes in tectonic or climatic forcing. The apparent acceleration of incision rates in several western North American river systems may occur for this reason, as the highest measured rates are obtained from the youngest dated terrace.

Because strath terraces imply long periods of lateral planation and nearly constant river elevation, they are inferred to reflect a time of balance between sediment supply and discharge such that neither net aggradation nor degradation occurs (i.e., $dQ_s/dx \approx 0$ and $dz/dt \approx 0$). This state is called a graded river (Bull, 1990; Mackin, 1937). Although lateral planation implies nearly constant elevation, our simulations suggest that these graded periods should not be taken to indicate steady forcing. All of the experiments produce terraces, often by abrupt onset of downcutting, even though the input is being changed continuously and even gradually (e.g., Figs. 4 and 9). The valley-widening periods are characterized by slightly positive divergences in sediment-transport capacity ($dQ_s/dx > 0$), where x increases downstream, reflecting a balance within a channel reach wherein all sediment delivered into a reach and the sediment generated by bedrock erosion are transported downstream. This balance keeps the sediment thickness on a forming strath relatively constant. The transition to downcutting occurs as dQ_s/dx increases, becoming more positive, allowing net sediment removal from the channel and enhanced access to the bedrock. Conversely, an increase in sediment supply, for instance, would push dQ_s/dx toward negative values (i.e., aggradation), switching the river from strath formation to valley filling. Fill terraces can be generated in such circumstances, which would be an interesting subject for a future modeling effort. A threshold exists between valley fill and strath-terrace formation that is dictated by the divergence in sediment-transport capacity, rather than implying distinctly different inputs and river response to those inputs. As a consequence, two identical river systems with only slight differences in the amplitude of fluvial inputs may produce two different records of the same event: a strath terrace in one and a fill terrace in the other.

The results point to several future modeling experiments. First, sediment should be delivered at tributary junctions as well as at the head of the trunk channel. Tributary inputs may influence the timing and magnitude of lateral planation in the down-valley direction, perhaps producing diachroneity along terrace profiles. Second, the quantity of sediment input and the ease with which the valley can be widened control the ability of the river to widen and form strath terraces, as opposed to aggrading to form fill terraces. Although we have focused on production of strath sequences here, as are seen in the Wind River system from which we deduced the parameter set used, it would also be interesting to explore the conditions leading to fill-terrace formation. Preliminary simulations suggest that less easily eroded valley walls promote the formation of significant fill terraces. Finally, although we have maintained a steady base-level fall here, a goal should be exploring how nonsteady base-level fall, with superimposed variations in river inputs, influences the resulting terrace sequence.

CONCLUSIONS

Starting with the conceptual hypotheses of Gilbert (1877) and Mackin (1937), we have created a physically based model that succeeds in producing distinct fluvial strath-terrace sequences in response to time-dependent inputs to the fluvial system. We have selected a physically based rule set and have set limits on inputs from available field evidence. Our numerical experiments indicate that strath terraces can be created through climatic modu-

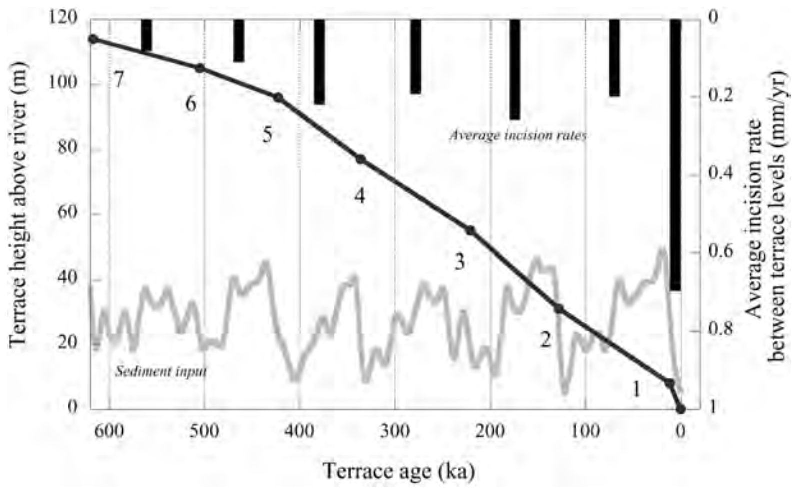

Figure 11. Graph of height of model terraces above the channel (as labeled on left *y*-axis) vs. terrace age produced by the tenfold sediment-supply simulation scaled by the oxygen isotope curve (sediment-supply-variation history shown by gray curve). The average incision rates between terrace levels calculated from terrace age and height are shown with bars (as labeled on the right *y*-axis). There is a general trend of increasing average incision rates toward the present (i.e., end of the simulation). A similar trend has been noted for several terrace sequences in western North America (see text for references), often based on terraces of ages similar to model terraces 1, 2, and 7 in this graph. Although such acceleration may be ascribed to increasing uplift rates and/or increasing climate intensity in the field, our results suggest this may only be an apparent acceleration given the unique time period that we are observing. In our simulation, this apparent acceleration occurs in spite of a constant tectonic forcing (i.e., base-level lowering). The high rate associated with the most recent terrace (level 1) is obtained because, since abandonment of this terrace level, the channel has been largely in a period of downcutting. Incision rates calculated between older terrace levels include long periods of little to no downcutting; hence, average rates tend to be lower solely because of the time sampled. Between the time of formation of terrace 7 and terrace 5 in our simulation, the stream spends an unusually long time forming terraces (i.e., little vertical incision) relative to the remainder of the simulation. Incorporating this period into the calculation of average incision between terrace level 7 and any terrace created after terrace 5 (like terrace 2) tends to produce a relatively low incision rate. Thus, our simulations explain why incision rates appear to increase, but do not require either increased uplift or enhanced climate severity.

lation of sediment load and water discharge delivered to a fluvial system. This modulation causes rates of lateral planation and river incision to vary through time, and this variation fosters creation of wide valley floors that are abandoned to form terraces during periods of more rapid incision. Terrace development and river response are strongly dependent on the amplitude of input fluctuations with time, and terrace width, maximum incision rates, and valley floor residence time increase as the input amplitude increases. Terrace width is controlled largely by the river's residence time at a particular elevation, and the width of climatically generated terraces is indicative of the duration and/or magnitude of climatic change.

Several aspects of model behavior, as yet undocumented in field settings, provide new ideas for investigation and interpretation of terrace sequences. The periods of lateral planation required to form the valley floors that become terraces may be very long. In our simulations, these surfaces were occupied for many tens of thousands of (model) years and a significant fraction of the overall river history. Deposition of gravel across and down a particular terrace tread is likely diachronous, with a potentially large variation in the timing of deposition. Given this variability, assigning a single numeric age to a terrace surface should be done with caution and with reference to the time within the terrace cycle (e.g., valley abandonment) to which the age corresponds. Although these periods of lateral planation with low incision rates are representative of the classic graded-river condition, achievement of grade in our simulations does not require that inputs be steady. Continuous lateral planation occurs in our simulations even though inputs are irregular and nonsteady. Terraces should not be interpreted to reflect a period of steady forcing, but instead a period of delicate balance achieved even as the external forcing on the fluvial system continues to change. The graded condition in the model fluvial systems reflects a delicate balance between incision and aggradation; destruction of this balance occurs in response to modest changes in fluvial inputs.

The simulations also indicate that the timing of terrace formation significantly lags the input cycles driving terrace formation. This observation is consistent with field documentation of a paraglacial period that follows major climate shifts during the transition from glacial to interglacial conditions. As in the field, our model fluvial systems require time to respond and adjust to changing inputs, particularly during the transition from periods of lateral planation to more rapid incision. Terrace formation occurs well after the onset of the input changes forcing their creation. Dates from terraces in the field should be expected to lag the onset of climate changes that affect fluvial inputs, such as the transition from glacial advance to retreat.

The incision-rate histories obtained from terrace sequences in extant fluvial systems may be distorted by current river incision. Many fluvial systems are now actively downcutting following an extended period of lateral planation during the last glaciation. The youngest terrace in these fluvial systems is often related to the Last Glacial Maximum, and in some cases abandonment of the Last Glacial Maximum valley floor may not yet be complete (e.g., Chadwick et al., 1997). Because of this active downcutting, incision rates obtained from the age and height of a Last Glacial Maximum terrace in these systems are averaged over a period of continual incision. Incision rates from older terraces, on the other hand, are averaged over both prolonged periods of lateral planation (i.e., incision rates near 0) and incision periods. As a consequence, incision rates calculated from the most recent terrace level would be high even if the long-term average incision rate is not changing. This apparent acceleration of incision rate has been observed in many fluvial systems, but should not necessarily be interpreted to indicate fluvial response to changes in external forcing, such as increasing rock-uplift rates.

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