STREAM PROFILE ADJUSTMENT TO CRUSTAL WARPING: NONLINEAR RESULTS FROM A SIMPLE MODEL[1]

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ABSTRACT

Three numerical experiments have been performed, modeling response of a greatly simplified stream system to initiation of crustal tilting and domal uplift. Model results show signs of nonlinear behavior in varying degrees. These signs include river profile adjustments that have forms quite different from the geometries of crustal movements, cases of channel erosion following deposition at the same location, and spatial transition points between opposing adjustments (i.e., erosion and deposition) that fail to coincide with boundaries or axes of crustal movement. Particular nonlinear effects wax or wane as the system adjusts to a new, dynamic equilibrium with the continuing crustal movement. The modeled features of stream adjustment are not in so extreme a class as the commonly discussed complex response of fluvial systems, but they do serve to suggest some caution in studies that employ stream profile information as an indicator of crustal movement.

INTRODUCTION

In a recent paper (Slingerland and Snow 1988) we presented numerical model results exploring the complex response of a fluvial system to base level lowering. Here we present results from a less sophisticated model of river profile adjustment following initiation of crustal tilting, and also domal uplift. These results are of interest for two reasons. First, they show that a very simple river system model can nevertheless exhibit several types of nonlinear behavior. Second, they suggest that actual river adjustments to crustal movement may be nonlinear in character, requiring an additional degree of caution in the interpretation of field data.

By *nonlinear behavior* in river profile adjustment we mean system response to an applied stress in which changes in river profile are not related to the stress by a simple additive function or proportionality, but rather show more complex forms. Similarly in time, channel variables along the profile shift from initial to new equilibrium values by paths more complex than simple asymptotic decay.

Complex response of the fluvial system (Schumm 1973) is most commonly described in terms of episodic sediment routing within tributaries and the trunk stream of a channel network. The model applied in the present study accounts for only a single channel with no allowance for tributary response to its intrenching or aggradation. For that reason, the nonlinear simulation results presented here are presumably in a different class from complex response, and also because they are not nearly so dramatic, nor do their causes require such delicate explanation.

A number of studies investigating active crustal warping make use of information related to river profile adjustments. For example, Volkov et al. (1967), Burnett and Schumm (1983), and Ouchi (1985) interpret irregularities in present-day river profiles as signs of vertical crustal movement. There are also numerous studies of crustal flexure based on evidence from fluvial terraces (Han 1985; Markewich 1985; Rockwell et al. 1988). If nonlinear aspects of river profile adjustment to crustal movement exist, then it is important to explore them by whatever means available so that overly rigid or unnecessarily crude interpretations of field data may be avoided.

In fact, almost all examples of nonlinear behavior discussed here are readily explained once they are observed. They represent secondary effects of the primary channel adjustments, which any qualitative, conceptual model would predict for given crustal movements. A key service given by quantitative models however, is their thorough, unbiased exploration of such secondary effects of fluvial response within the limitations of model completeness. In doing this they can add so-

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phistication to the conceptual models used to interpret field data.

MATHEMATICAL MODEL AND SIMULATED STREAM CHARACTERISTICS

The mathematical model applied here describes nonuniform, unsteady flow and sediment routing in a channel with erodable, alluvial bed. Controlling equations are:

conservation of fluid momentum:

$$W\frac{\partial[u(h-b)]}{\partial t}+\frac{\partial[u^2W(h-b)]}{\partial x}+gW(h-b)\frac{\partial}{\partial x}=Wgu\frac{|u|}{C^2} \tag{1}$$

continuity of water:

$$W\frac{\partial(h-b)}{\partial t}+\frac{\partial[uW(h-b)]}{\partial x}=Q_i \tag{2}$$

continuity of sediment:

$$\alpha W\frac{\partial b}{\partial t}+\frac{\partial[Wq_s]}{\partial x}=Q_{si}+\alpha WY \tag{3}$$

in which t is time, x is distance downstream, W is channel width, u is flow velocity, h is water surface elevation, b is bed surface elevation, g is acceleration of gravity, C is Chezy flow resistance coefficient, Q_i is lateral inflow of water in units of discharge per unit length of channel, Q_{si} is volumetric lateral inflow of sediment in the same units, alpha is one minus the porosity of channel sediments, q_s is volumetric sediment transport per unit width, and Y is local crustal uplift (or downdrop) rate. Beyond common assumptions for such equations, these in particular involve assumptions that flow is hydraulically wide and that lateral inflows carry with them no downstream-directed momentum. Sediment transport is calculated here by means of the Engelund bedload equation (Engelund and Fredsøe 1976).

The model equations are solved numerically in the FORTRAN program LPMOD, which employs an implicit finite-difference solution method (Fread 1978) adapted for fully coupled sediment routing and modified to a mixed-difference formulation (Snow 1988). In this study, externally imposed water and sediment inflows to the stream channel are held constant through time, so that a simpler, known discharge equation set would suffice. LPMOD is overdesigned for this application, but it adapts well to constant-discharge situations, giving solutions over the required long time steps at quite low computational expense, and allows such modifications as the inclusion here of an uplift term to be made with ease. Alongstream data points are spaced evenly at a 5 km interval for the model runs.

This model is a grossly incomplete representation of natural fluvial systems in many ways, including the following model characteristics:

1. Channel cross-section geometry does not alter with erosion or aggradation.
2. Erosional lowering of the modeled channel produces no additional influx of sediment from tributaries, channel banks, or valley sides.
3. Aggradation only operates within the limits of the channel width, so that no sediment volume is dispersed as general valley fill.
4. A straight channel planform (sinuosity 1.0) is assumed. Secondary flow directions, lateral channel erosion, and temporal changes in sinuosity are not considered.
5. A single diameter value represents local sediment characteristics, with no accounting for size sorting or stream bed armoring.
6. An infinite depth of noncohesive, alluvial material is assumed. Sediment transport rate is never less than the flow's capacity.
7. Hydraulic conditions and sediment transport are taken to have no effect on flow resistance, as expressed by a constant Chezy C value.

On the whole, this model is more like a numerical flume than a natural stream.

The chief purpose of this paper is to show that even such a simplified system as this can have complicated behavior. Many fluvial process relationships, with various threshold conditions and response times, have been ignored here. Were all these further complexities to be added to the model, it seems reasonable to assume they would add to, rather than subtract from, the nonlinearity of system behavior. This assumption of increasing complexity of behavior with increasing degrees of freedom is not universally accepted.

TABLE 1

CHARACTERISTICS OF MODELED STREAM

Value	Controlling Relationship and/or Range of Values (x = 0–200 km)
Discharge, Q (m^3s^{-1}) = ΣQ_i	$Q = 0.5\ m^3s^{-1} + 0.0475\ m^3s^{-1}km^{-1}x$ (0.5–10)
Sediment Discharge, Q_s (m^3s^{-1})[a] = ΣQ_{si}	$Q_s = 0.00003\ m^3s^{-1}km^{-1}x$ (0.0–0.006)
Sediment Diameter d (mm)	$d = 6\ mm\ e^{-.0055km^{-1}x}$ (6.0–2.0)
Channel Width W (m)	$W = 8\ m^{-.5}s^{.5}Q^{.5}$ (5.7–25)
Channel Gradient[a]	(0.0039–0.0009)
Depth of Flow (m)[a]	(0.14–0.66)
Flow Velocity (ms^{-1})[a]	(~0.6)
Flow Resistance (SI units)	$C = 25$
(1 − Porosity) of Channel Sediment	$\alpha = 0.7$

[a] Denotes initial condition, before crustal warping.

Some would argue that a system with more feedback loops and large number of elements is likely to be more stable to perturbations. This is part of an ongoing debate in ecology today, with ramifications for many analogous systems including sedimentary systems (Slingerland 1989). In later sections of this paper we offer qualitative arguments that addition of at least some degrees of freedom to the model would preserve or enhance the nonlinearity of channel profile response to the simple types of crustal movements examined.

A single modeled stream reach, 200 km in length, is used for all numerical experiments in this study. In addition, the stream modeled here is generally similar to the trunk stream modeled in the earlier study of complex response (Slingerland and Snow 1988) and is almost identical to the modeled stream responding to base level lowering in another paper (Snow 1988). In general terms, it is a weakly effluent, granule-bed stream.

Given the model simplicity described above, we consider it appropriate to include only first-order approximations to alongstream changes in stream characteristics. The smooth, linear alongstream increases in water and sediment discharges (table 1) give a simple background for interpretation of model results. The zero value of sediment transport paired with a nontrivial water flow of 0.5 m^3s^{-1} at the upstream end ($x = 0$) is not very realistic, except as a representation of outflow from a lake. However, it maintains similarity with model runs treating river response to simple base level drops (Snow 1988) should a comparison of results be of interest. The imposed exponential decrease in grain size alongstream is included as a widely-supported empirical relation, with no assumption as to whether its cause is comminution or sorting.

Upstream discharge and downstream base level provide adequate hydraulic boundary conditions for the model. Sediment continuity at the downstream end is calculated by a backward-difference form of equation (3), and bed elevation change through time at the upstream end, a difficult boundary value to formulate realistically, is made equivalent to the bed elevation change at the nearest downstream data point. The mixed-difference solution method employed (Snow 1988) allows such a crude upstream boundary condition to be imposed with negligible effect on adjacent downstream solution values. This has been confirmed in a specific case mentioned in the section below. The program LPMOD has been run with all of the above input data to produce a river profile at equilibrium with conditions of no crustal movement. Ranges of gradient and flow characteristics for this profile are given in table 1. This is the initial condition for each of the profile-adjustment runs described below.

PROFILE ADJUSTMENT TO CRUSTAL MOVEMENTS

Three simulations have been run, modeling river profile response to crustal tilting and warping. In the first two runs, crustal movement only affects stream conditions by changing local channel gradients; in the third, relative change in base level also occurs. In each run the crustal movement initiates abruptly at the run beginning and proceeds at constant rate thereafter. No lateral channel movement is considered, in part because we assume that river flow direction is across, rather than along, contours of uplift rate.

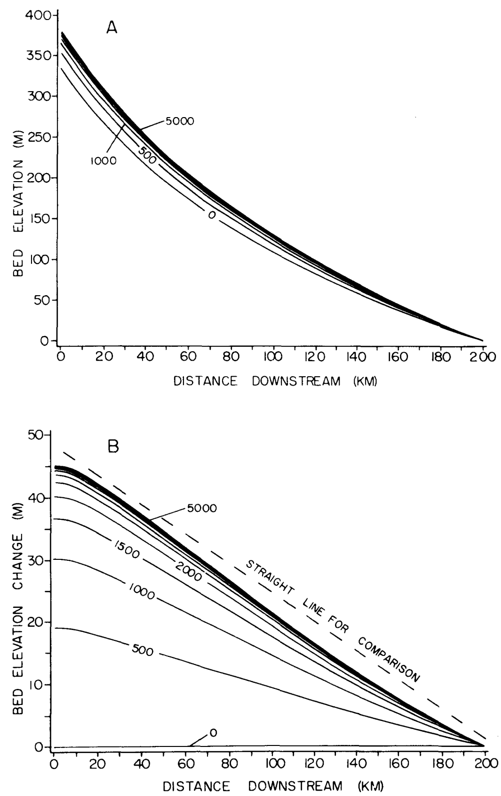


FIG. 1.—Stream profile adjustment at various times (years of model time) for modeled stream response to tilting about a downstream axis. Crustal uplift rate at the upstream end is 5 cm-yr^{-1}. *A*—Plot of stream bed elevation (longitudinal profile). *B*—Net bed elevation change from initial state.

Tilting about Downstream Axis.—In the first run, a simple tilt is imposed at a rate of 0.25 microradians per year, with the downstream end of the modeled stream reach remaining unchanged in elevation while the upstream end uplifts at a rate of 5 cm/yr. It would be quite a coincidence for a stream to meet base level at just the location of an axis of tilting, as this implies. However, consideration of relative rather than absolute elevation change allows this case to fit any situation in which base level elevation is dependent on the crustal movement, as it would be if the local base level control were a highly resistant rock lip or a major trunk stream running along contours of crustal uplift.

Results of the run are shown as adjustments of river profile in figure 1*A*. The increasing slope all along the channel produces erosion that accelerates as tilting progresses, until a stable longitudinal profile of dynamic equilibrium is developed. In first approximation, the elevation difference between initial and final equilibrium profiles is directly proportional to the local amount of tilt-related uplift. Yet in detail (fig. 1*B*), the ultimate profile of bed elevation change is not linear but weakly convexo-concave. Note that the short, level segments of profiles within 5 km of the upstream boundary in figure 1*B*, and in similar figures elsewhere in this paper, are expressions of the artificial boundary condition that equates bed elevation changes at the two most upstream data points. Ignoring these, it is still apparent that the net change in stream profile does not completely mirror the simple geometry of tilting.

Possibly the first explanation that comes to mind is that this second-order deviation of the profile form, especially the relative convexity near the upstream end, is a spurious result of numerical errors due to the boundary condition. To test the sensitivity of the model results to this imposed condition, we have duplicated the model run substituting two even more inaccurate conditions. In the first alternate run, the upstream bed elevation value is held constant, despite continuing crustal uplift downstream. The data point quickly becomes the site of a deep pool. For the second run, the upstream bed elevation value passively rises with crustal uplift, allowing no local erosion and gradually creating an anomalously steep channel gradient to the next downstream point. After 5000 yr of model time, computed bed elevation values for the original and two alternate runs are compared: 250 m difference at the most upstream point, <1 cm difference at the next point 5 km downstream, <1 mm 100 km downstream. The boundary condition has negligible effect on the general profile form.

The cause of the nonlinearity in profile adjustment rests elsewhere—in the newly imposed requirements for sediment transport.

The volume of channel sediment that must be eroded to balance uplift is greater in upstream locales than downstream, although the difference is not directly proportional to uplift rate as it also depends on local channel width. Downstream channel segments must adjust to transport all sediments being eroded at points upstream, but also have had comparatively greater sediment loads and transport capacities in their initial equilibrium states. Experiments by Snow and Slingerland (1987) confirm that strong relative increase in volume of sediment load along a stream can result in a convex longitudinal profile or at least a significantly reduced profile concavity. Analysis of the present model output data indicates that the local alongstream increase in required sediment transport, expressed as a percentage of net local sediment transport, is greater for the case with uplift than the case without in the upstream half of the river length, but is less for the uplift than the non-uplift case in the downstream half; hence the reduction in concavity upstream and increase in concavity downstream for the dynamic equilibrium profile.

The above interpretation suggests that an even stronger nonlinearity of dynamic equilibrium profile would result if tributary and hillslope response were considered, because the greatest additional sediment influx from these would occur upstream. In general, it is not likely that changes in equilibrium profile form will directly express more than the gross geometry of crustal tilting.

In this model run, following initiation of uplift there is a time lag of several thousand years before a profile of dynamic equilibrium develops. While the magnitude of such a time lag would presumably depend much on the characteristics of the stream system studied, this model run nevertheless highlights the possibility that a natural stream studied in the field may not have reached a condition in which channel erosion rates approximate current uplift rates. For example, during the first 500 yr of this model run about 80% of the crustal uplift is expressed in profile change, with the remainder being balanced by vertical channel erosion. During this period uplift rate would be strongly underestimated by a calculation from river terraces, which express erosion rate.

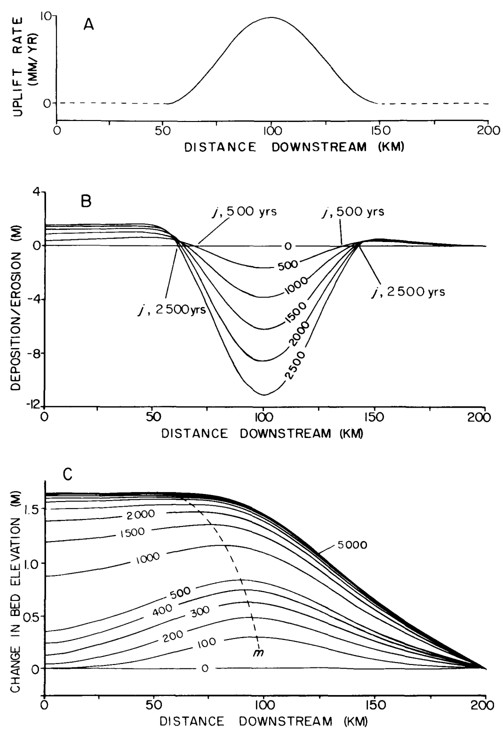


FIG. 2.—Modeled stream response to domal uplift centered 100 km downstream. *A*—Uplift rates imposed for various alongstream locations. *B*—Net channel deposition (+) and erosion (−) at various times (years of model time). *C*—Net change in stream profile from initial state, which includes effects of both crustal movement and deposition/erosion. Initial longitudinal profile for this model run is the same as the initial profile shown in figure 1*A*. Symbols *m* and *j* are defined in figure 3.

Domal Uplift.—In the second model run, a domal or anticlinal uplift is imposed along the stream course. The zone of uplift is limited to the central 100 km of the modeled reach, and within that zone uplift rate varies as a sine function (fig. 2*A*). Maximum uplift rate is 1 cm-yr^{-1}.

For this experiment, and the one to follow, it is important to view the results within two different frames of reference. We can view channel adjustments in terms of deposition and erosion of channel sediments (fig. 2*B*). These are the kinds of changes that would be recorded by measurements from local fixed points on the uplifting land surface (streambed elevation change minus land elevation change due to uplift). This is also the type of information that can be recorded in zones of net erosion by river terraces. Alternatively, the channel adjustments can be viewed as they were for the first model run, in terms of longitudinal profile modification; that is, bed elevation change as monitored from some universal datum (fig. 2*C*).

Erosion in the central portion of the stream profile is accompanied by deposition downstream that increases local channel slope and allows transport of the additional, erosion-related sediment load (fig. 2*B*). Deposition also occurs in the upstream zone as local base level rises with bed elevation in the central zone (fig. 2*C*). Similar zones of aggradation upstream and downstream of a local uplift have been previously observed in flume experiments (Ouchi 1985). In the present numerical model results, the upstream channel filling initially has a wedge shape in approximate symmetry to downstream channel filling but later develops to an equal depth of aggradation extending to the upstream end. Deposition rates are of the same order as erosion rates for the first few hundred years of model time, but thereafter, depositional areas reach general stability while continuous erosion elsewhere balances continuous uplift. The central zone of net erosion (*ne*, fig. 3) gradually expands in time, approaching the spatial limits of uplift activity and truncating earlier channel fill.

The way crustal movement is expressed in stream profile change (fig. 2*C*) depends strongly on the stage of profile adjustment. At a model time of 100 yr, bed elevation change mimics fairly well the general form of uplift, but extends farther upstream and downstream than the limits of actual crustal warping. As fluvial adjustment progresses, the asymmetric nature of deposition becomes evident, and the elevation change profile in the zone of uplift evolves to a broadly convex profile of increased channel gradient. As a result, the maximum point of net profile change (*m*, also in fig. 3) migrates several tens of kilometers upstream of the location of maximum uplift.

If tributary and valley side response were to be included in this run, the major effect would be a heightened amount of uplift-generated sediment production in the central stream course. This would require a greater magnitude of profile adjustment throughout

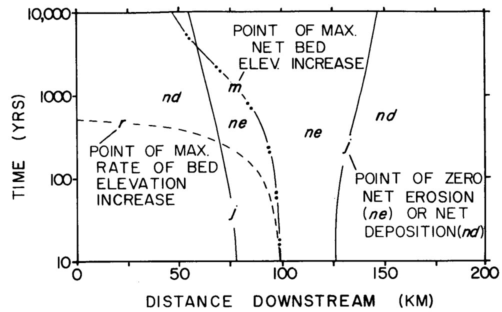


FIG. 3.—Shifting of various transition points and maxima of profile adjustment along river course through time, in modeled stream response to domal uplift centered 100 km downstream. Note logarithmic scale for time. Locations are shown for points of zero net deposition/erosion (*j*, fig. 2*B*), zones of net deposition (*nd*) and net erosion (*ne*), point of greatest net rise in bed elevation from the initial state (*m*, fig. 2*C*), and point of greatest momentary rate of bed elevation increase (*r*).

the modeled reach, but it would not alter the general form of adjustment. The simplicity of this model does not allow for another means of river response to domal uplifts: change in channel sinuosity, as discussed by Adams (1980) and Burnett and Schumm (1983). However, some classes of rivers lack that degree of freedom.

In each of the two numerical experiments so far described, the stream channel responds to uplift by developing a new dynamic equilibrium profile characterized by local increases in convexity. Several field studies have documented instances of anomalous river profile convexity in zones of crustal uplift (Russ 1982; Burnett and Schumm 1983; Ouchi 1985; Rhea 1988). The model results presented here serve to emphasize that there are two ways to interpret these convexities; they need not be taken as symptoms of disequilibrium, of a fluvial system insufficiently adjusted to tectonic stress, such as is exemplified by the 100-yr profile in figure 2*C*. They could well be features of dynamic equilibrium adjusted to a fairly continuous stress, as is the ultimate profile in figure 2*C*. In either case (especially the latter) it is important to note that the profile "disruption" may extend spatially far beyond the area of crustal movement, and that the crests of profile convexities may be found well upstream of sites of maximum uplift.

Tilting about Central Axis.—In this final run we again model effects of simple tilting but set the tilt axis midway along the stream course ($x = 100$ km), representing a case in which base level is independent of crustal movement. The tilt rate is 0.05 microradians per year, producing uplift of 5 mm/yr at the upstream end and an equivalent rate of drop below base level at the downstream end.

The resulting profiles of erosion and deposition (fig. 4*A*) show that the onset of deposition near the downstream end, a result of base level control, precedes the onset of significant erosion resulting from increase in channel gradient (see profiles at 100, 500 yr). Later in the run, profiles of deposition and erosion become approximately linear, and the points of zero net erosion/deposition (*j*) and zero erosion/deposition rate (*k*, fig. 4*D*) shift to locations approximating the position of the true tilt axis (*a*).

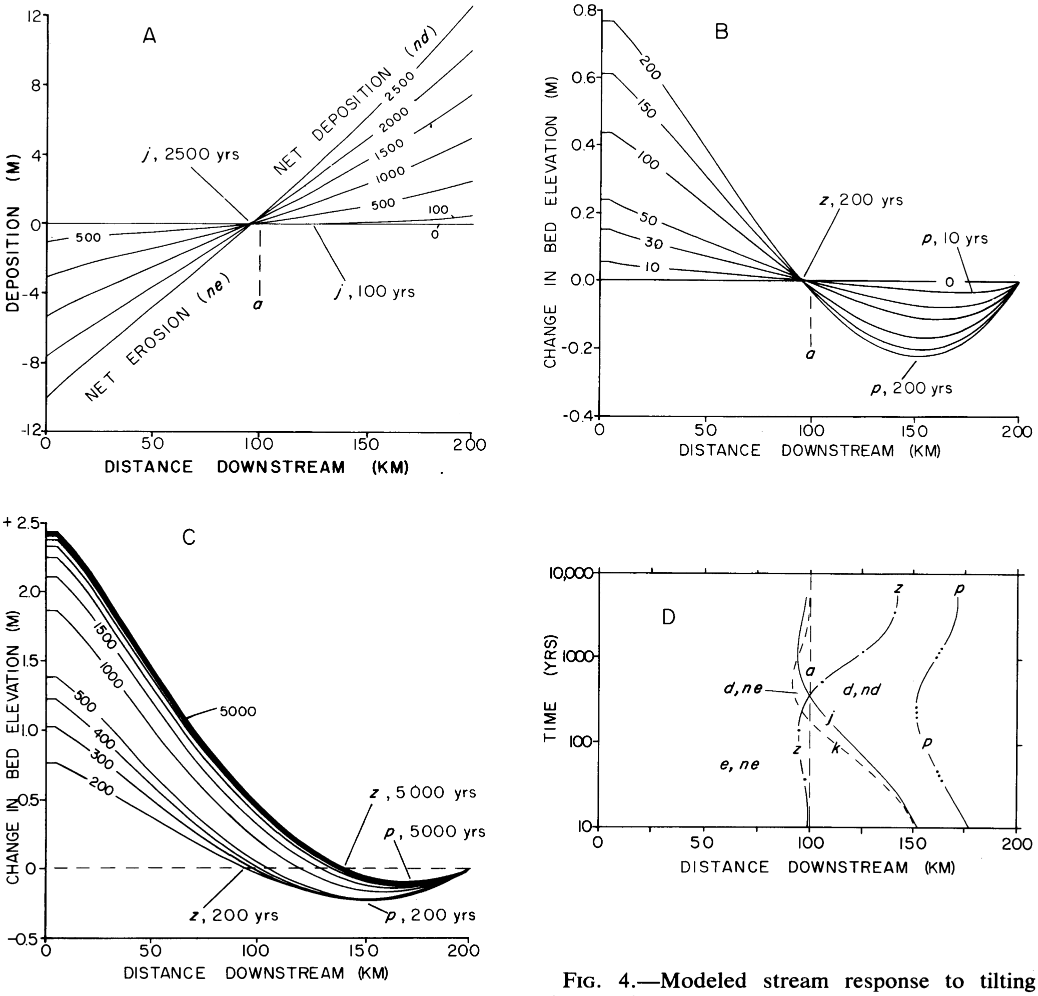
A more complex behavior is shown by changes in the stream profile. During the first 200 yr of model time (fig. 4*B*), bed elevation is lowered in the zone downstream of the tilt axis, with the location of maximum bed elevation loss (*p*) shifting upsteam within that zone. There is no appreciable loss in bed elevation at the downstream end where the river meets base level, although only water surface elevation is fixed there. However, any loss of bed elevation there directly results in an increase of flow depth, loss of velocity, and sediment deposition that rapidly matches the imposed crustal lowering rate. During the period of time shown in figure 4*B*, the point along the stream with zero net change in bed elevation (*z*) remains close to the tilt axis position but shifts about 5 km upstream (fig. 4*D*). In the time that follows (fig. 4*C*), downstream bed elevation loss is first limited, then reversed, by profile gradient requirements to transport sediments being continuously eroded upstream. The point of zero net change in bed elevation (*z*) migrates far downstream of the true tilt axis, and the final, dynamic equilibrium profile is characterized by slightly increased concavity in all but its most upstream sections.

Nonlinearity appears as a first-order feature of profile elevation change both in this model run and the previous run, although the net profile adjustment in each case is small compared to the whole profile relief (about 350 m). Many of the nonlinearities in deposition and erosion, which might be recorded in terraces, are most strong early in the history of system evolution. Could the various nonlinear effects be more prominent and lasting than in the example runs given here, provided different stream characteristics and crustal warping conditions? It is reasonable to accept this possibility.

A modified version of the last run, including a five-fold increase in tilting rate, results in the same changes shown in figure 4 increased by a factor of five. Within this range, at least, profile change scales proportionately with rate of crustal warping. We have not explored mathematically the ways the effects change in magnitude or type with changes in river size, slope, and so on. However, it is notable that the example river conditions used in this study were not specifically chosen to maximize nonlinearity or magnitude of profile adjustment, and assuming that the different classes of streams will vary considerably in this way, a significant subset of these classes is likely to exhibit stronger response.

This model run exhibits reversal of bed elevation change, alongstream shifts in transition points and maxima of profile adjustment through time, and strong temporal differences in rate and direction of those shifts. These all result from a difference in lag times for river adjustments responding to the two separate stresses imposed on the stream system: increase in stream gradient from differential uplift, and relative base level change. This basic nonequivalence of lag times can reasonably be expected to persist even if such factors as estuary fill requirements, channel armoring, and tributary response to main channel adjustment were taken into account.

Han (1985) shows schematic examples of rivers in China that cross adjacent upwarp and downwarp zones and have terrace intersections that shift alongstream with terrace level. Han interprets these shifts in terrace convergence as signs of tectonic migration of the upwarp center. An alternative explanation is suggested by the results of the second and third model runs presented here (fig. 2*B*, 4*A*), although on what is probably a smaller



SYMBOLS

a - AXIS OF CRUSTAL TILTING
d - ZONE OF ACTIVE DEPOSITION
e - ZONE OF ACTIVE EROSION
k - POINT OF ZERO ACTIVE DEPOSITION/EROSION
nd - ZONE OF NET DEPOSITION
ne - ZONE OF NET EROSION
j - POINT OF ZERO NET DEPOSITION/EROSION
z - POINT OF ZERO NET CHANGE IN STREAM BED ELEVATION
p - POINT OF MAXIMUM NET LOSS IN STREAM BED ELEVATION

FIG. 4.—Modeled stream response to tilting about axis centered along stream course. Uplift and downdrop rates at upstream and downstream ends are 5 $mm\text{-}yr^{-1}$. Locations are marked for the axis of crustal tilting, zones of deposition and erosion, and various transition points and maxima of profile adjustment. *A*—Net channel deposition (+) and erosion (−) at various times (years of model time). *B*—First 200 yr of stream profile change. Initial longitudinal profile is the same as the initial profile shown in figure 1*A*. *C*—Remaining profile adjustment after 200 yrs model time, shown as net bed elevation change from initial state. *D*—Shifting of transition points and maxima of profile adjustment along river course, compared to location of true tilting axis. Note logarithmic scale for time.

spatial and temporal scale. According to this view, the shifts may instead show dynamic response of the fluvial system to recently initiated or intermittent, but positionally stable, crustal flexure.

CONCLUSION

The three numerical experiments described in this paper represent the response of a highly simplified stream system to uncomplicated crustal movements, yet they provide examples of several types of nonlinear behavior: (1) river profile modifications, whether in the process of adjustment or at dynamic equilibrium, that have forms divergent from the geometries of crustal movements, (2) local reversals of deposition/erosion within the history of adjustment to continuing crustal movement, (3) significant lag times between inception of crustal movements and notable channel erosion responding to the movements, and (4) transition points of deposition/erosion and riverbed elevation change that do not consistently mark locations of true axes of crustal movement, and that along with profile adjustment maxima shift significant distances alongstream while crustal movement remains constant.

We affirm that these occurrences fall short of the complex stream adjustment typically given the name *complex response,* and that the stream model employed here greatly underrepresents the true complexity of the fluvial system. We consider the nonlinear effects discussed here to be of interest precisely because of the simplicity of the system involved.

In addition, the general types of nonlinear behavior observed here do not appear to be directly linked to the simplifying assumptions made. That is, we have considered some real fluvial system characteristics that would be likely to modify significantly sediment routing as modeled here, and we believe it is a reasonable assumption that the complexities observed here would be modified, but not negated as general types, in a more complete system model.

For the above reason, we believe these results are adequate to suggest some caution in studies that seek to interpret fluvial terraces and other field records of stream profile adjustment as surrogate records of continuing crustal movement, particularly the detailed geometry of such crustal movement.

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REFERENCES CITED

ADAMS, J., 1980, Active tilting of the United States midcontinent: geodetic and geomorphic evidence: Geology, v. 8, p. 442–446.

BURNETT, A. W., and SCHUMM, S. A., 1983, Alluvial-river response to neotectonic deformation in Louisiana and Mississippi: Science, v. 222, p. 49–50.

ENGELUND, F., and FREDSØE, J., 1976, A sediment transport model for straight alluvial channels: Nordic Hyd., v. 7, p. 293–306.

FREAD, D. L., 1978, National Weather Service operational dynamic wave model: 26th Ann. Hydraulics Div. Specialty Conf. Proc. (College Park, Md.), ASCE, p. 455–464.

HAN, M., 1985, Tectonic geomorphology and its application to earthquake prediction in China, *in* MORISAWA, M., and HACK, J. T., eds., Tectonic Geomorphology: Boston, Allen & Unwin, p. 367–386.

MARKEWICH, H. W., 1985, Geomorphic evidence for Pliocene-Pleistocene uplift in the area of the Cape Fear Arch, North Carolina, *in* MORISAWA, M., and HACK, J. T., eds., Tectonic Geomorphology: Boston, Allen & Unwin, p. 279–297.

OUCHI, S., 1985, Response of alluvial rivers to slow active tectonic movement: Geol. Soc. America Bull., v. 96, p. 504–515.

RHEA, S., 1988, How can river valley morphology help to identify areas of differential uplift in the western Oregon Coast Range?: Geol. Soc. America Abs. with Prog., v. 20, p. 307–308.

ROCKWELL, T. K.; KELLER, E. A.; and DEMBROFF, G. R., 1988, Quaternary rate of folding of the Ventura Avenue anticline, western Transverse Ranges, Southern California: Geol. Soc. America Bull., v. 100, p. 850–858.

RUSS, D. P., 1982, Style and significance of surface deformation in the vicinity of New Madrid, Missouri: U.S. Geol. Survey Prof. Paper 1236, p. 95–114.

SCHUMM, S. A., 1973, Geomorphic thresholds and complex response of drainage systems, *in* MORISAWA, M., ed., Fluvial Geomorphology: Binghamton, NY, State Univ. New York Pub. in Geomorphology, p. 299–310.

SLINGERLAND, R. L., 1989, Predictability and chaos in quantitative dynamic stratigraphy, *in* CROSS, T. A., ed., Quantitative Dynamic Stratigraphy: Englewood Cliffs, NJ, Prentice Hall, p. 45–53.

———, and SNOW, R. S., 1988, Stability analysis of a rejuvenated fluvial system: Zeitschrift für Geomorph., Suppl.-Bd. 67, p. 93–102.

SNOW, R. S., 1988, A flexible program that allows long-term modeling of river profile adjustments: Computer Oriented Geol. Soc. Computer Contrib., v. 4, p. 80–107.

———, and SLINGERLAND, R. L., 1987, Mathematical modeling of graded river profiles: Jour. Geology, v. 95, p. 15–33.

VOLKOV, N. G.; SOKOLOVSKY, I. L.; and SUBBOTIN, A. I., 1967, Effect of recent crustal movements on the shape of longitudinal profiles and water levels in rivers, *in* Symposium on River Mechanics (Bern): Int. Union Geodesy Geophys. Pub. 75, p. 105–116.