How do pediments form?: A numerical modeling investigation with comparison to pediments in southern Arizona, USA

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ABSTRACT

Pediments are gently sloping, low-relief bedrock erosional surfaces at the bases of mountain ranges. Pediments tend to form more readily in arid climates and in weathering-resistant lithologies, but the processes responsible for pediment formation are still not widely understood after more than a century of debate. In this paper, I investigate the behavior of a coupled numerical model for the evolution of mountain ranges and their adjacent piedmonts that includes bedrock erosion in channels, soil production and erosion on hillslopes, and the flexural-isostatic response of the lithosphere to erosional unloading. For relatively small values of the flexural parameter, erosion of the mountain range leads to sufficient flexural-isostatic tilting of the adjacent piedmont that a suballuvial bedrock bench is exhumed to form an erosional surface on the piedmont. In addition, slope retreat at the mountain front and subsequent tilting of the abandoned surface can contribute to pediment formation by lengthening the pediment in the upslope direction. The rate of erosion on the piedmont must also be greater than or equal to the rate of soil production, thereby creating an erosional surface that has, at most, a thin veneer of soil or regolith. The rate of soil production depends primarily on climate and lithology, with lower soil production rates associated with more arid climates and more resistant lithologies. The model predictions are compared to morphometric analyses of pediments in the southwestern United States and to the detailed morphology of two classic pediments in southern Arizona.

INTRODUCTION

A pediment is a gently sloping, low-relief bedrock erosional surface at the base of a mountain range. In the United States, pediments are most commonly formed on weathering-resistant rocks in the Mojave and Sonoran Deserts of southeastern California and southern Arizona. Figure 1 illustrates an example of a pediment on the west side of the Santa Catalina Mountains near Tucson, Arizona. The pediment surface, in the foreground, is composed of the same granite that makes up the steeper portions of the Santa Catalina Mountains to the east. Given that the range and pediment are composed of the same rock type, what processes account for the abrupt slope break between the range and pediment? When the geomorphologists of the early twentieth century first came to the southwestern United States, pediments were among the most striking and puzzling features of the landscape. If fluvial channels respond to tectonic uplift by incising, how could the channels draining these ranges have maintained such low-relief surfaces as they downcut? If the pediment is formed primarily by slope retreat, how could the pediment be devoid of alluvium and regolith despite its gentle slope?

Pediment studies have a long history in the geomorphic literature. Pediments can be broadly classified into two main types: planation surfaces formed on less resistant bedrock in contact with steeper, more resistant bedrock (examples of which were described by, e.g., Gilbert [1877] and Miller [1950], and “rock pediments” composed of the same lithology, typically granite, granodiorite, or quartz monzonite, as that of the adjacent mountain range [Oberlander, 1997]). In this paper, I focus on the formation of rock pediments (herein referred to simply as pediments) because these are the more enigmatic of the two types.

Figure 1. Aerial photographs (south looking) of the Catalina pediment near Catalina, Arizona (northwest of Tucson). The close-up view (B) illustrates that the pediment is dissected with bedrock channels that grade into the steeper bedrock channels of the Santa Catalina Mountains. The pediment is bounded on its west side by the Pirate Fault. Aerial Photography by Peter L. Kresan ©1990.
The puzzle of pediment formation can perhaps best be understood by considering why pediments do not form along most mountain fronts. Pediments can fail to form for at least two reasons. First, most pediments in the Basin and Range are uniformly covered with alluvial fan deposits. In areas of rapid late Cenozoic uplift, rates of alluvial fan deposition are often sufficiently high that alluvial fans and bajadas uniformly cover the piedmont. The fact that areas of the Basin and Range characterized by neotectonic activity tend to have continuous alluvial cover suggests that pediment formation is a post-tectonic process (Dohrenwend, 1994). Second, even if bedrock is exhumed from beneath an initial cover of alluvium on the piedmont, the soil or regolith on that surface may be sufficiently thick to preclude classifying the surface as a pediment. For example, the piedmont surface of the eastern United States is a regionally extensive, low-relief bedrock surface bounded by the Blue Ridge and Appalachian Mountains to the west and the Atlantic Coastal Plain to the east. As such, the piedmont surface meets one of the criteria for pediments: it is a gently sloping, low-relief erosional piedmont surface. Flexural-isostatic tilting has been invoked as a means to maintain the relief of the piedmont surface despite the great age of the Appalachian orogeny (Pavich, 1989). This surface cannot be considered a pediment, however, because it is covered with regolith that is locally over 30 m thick (Crickmay, 1935). The great thickness of this regolith cover is due, in part, to the relatively humid climate of the eastern United States. In this paper, I argue that under conditions of sufficient erosional unloading and low fluvial rigidity, flexural-isostatic tilting can tilt piedmonts to the point of developing an erosional bedrock surface, by exhumation and/or by slope retreat, that is steep enough to prevent alluvial deposition. In addition, however, pediment formation requires a sufficiently arid climate and/or a weathering-resistant lithology so that the rate of erosion on that surface is greater than the rate of soil formation, thereby precluding the accumulation of a thick mantle of regolith over time.

Four principal processes have been invoked for pediment formation: lateral corrosion, sheetflow erosion, subsurface weathering, and exhumation and/or slope retreat. In the lateral corrosion model proposed by Gilbert (1877), Blackwelder (1931), and Johnson (1931, 1932), channels draining the mountains are thought to erode laterally as they downcut, thus maintaining a planar surface. Later studies concluded that the corrosion model is unlikely to work due to the fact that channels would have to downcut with uniform efficacy within a 180° swath in order to preserve the linear mountain fronts adjacent to many pediments (Davis, 1930a, 1930b; Rich, 1935; Ruxton, 1958; Ruxton and Berry, 1961; Mabbutt, 1966; Warneke and Stone, 1966; Hadley, 1967). Models that invoke sheetflood- ing to form pediment surfaces (e.g., McGee, 1897; King, 1949) are also generally regarded as unlikely to work (Cooke et al., 1993). The observation that pediments tend to form in granitic rocks has led to the hypothesis that pediments form by subsurface weathering (granitic rocks are particularly susceptible to this process) (Mabbutt, 1966). Pediment formation by slope retreat and exhumation of a suballuvial bedrock bench was proposed by Paige (1912), Lawson (1915), and more recently by Cooke (1970). Lawson envisioned that mountain fronts erode primarily by slope retreat, leaving behind an alluvium-mantled bedrock surface. Pediments are then exhumed from beneath the alluvium by tilting or doming. Cooke (1970) favored the exhumation model because it best explained the morphometric correlations he observed between pediments and upstream drainage basins in the western Mojave Desert. No clear mechanism has been proposed for causing the uplift that leads to bedrock exhumation in this model, however. Uplift cannot simply occur by reactivation of range-bounding faults, because that would cause incision of any pediment formed upslope from the fault and deposition on any pediment formed downstream of the fault. All of these models remain conceptual, and none has gained widespread acceptance for even a subset of pediments found in nature.

Strudley et al. (2006; Strudley and Murray, 2007) were the first to model pediment formation numerically. Numerical modeling is an attractive approach to this problem because it provides a means of investigating the coupled evolution of geomorphic process zones that is required to test the conceptual models that have been proposed over the past century. Strudley et al. (2006; Strudley and Murray, 2007) applied the soil production function concept of Heimsath et al. (1997, 1999) to argue that pediments form, in part, due to a negative feedback between the soil and regolith thickness and the rate of bedrock weathering. In this feedback, an increase in the thickness of regolith leads to a decrease in the bedrock weathering rate, which, under conditions of constant base-level lowering, tends to steepen the piedmont and result in soil and regolith removal. In their model, this feedback mechanism results in an equilibrium soil cover on the piedmont in which the rate of bedrock weathering beneath the regolith matches the rate of base-level lowering. Under certain conditions, this equilibrium soil thickness can be relatively thin, leading to a pediment. Strudley et al. also showed that if a “humped” soil production function is used, tors and inselbergs may form. The Strudley et al. model represents an important advance in our understanding of pediments, but more work is needed. One drawback of the Strudley et al. approach is that the rate of rock uplift of the mountain range adjacent to a pediment (or, equivalently, the rate of base-level lowering of the downstream model boundary), is assumed to be a prescribed value independent of other processes operating in the model. In most parts of the Mojave and Sonoran Deserts, however, active tectonic uplift has not occurred for millions or tens of millions of years. Hence, rock uplift in these regions occurs primarily as the flexural-isostatic response to erosion in the ranges. A second drawback of the Strudley et al. approach is that it assumes that all channels in the model are alluvial channels. In fact, most channels on pediments and their adjacent mountain ranges are bedrock channels.

Figure 2 illustrates the conceptual model of this paper. In the early stages of pediment development, tectonic extension forms a semi-periodic sequence of basins and ranges. Hillslope and channel erosion in these ranges occurs in response to the relief production associated with extension. Erosion reduces the topographic load, triggering flexural-isostatic rebound. If the flexural parameter α (a function of the flexural rigidity of the lithosphere and the density contrast between the crust and the mantle) is relatively large relative to the spacing between ranges, isostatic rebound will be uniformly distributed across basins and ranges, resulting in regional rock uplift with little or no tilting. In the absence of tilting, alluvium and/or colluvium will be deposited on the piedmont, thus preventing pediment formation. Conversely, if the value of α is small relative to the spacing between ranges, erosion will result in concentrated isostatic rebound beneath the ranges. As a result, the piedmonts at the flanks of each range will be tilted, causing a bedrock surface to be exhumed, if sufficient tilting occurs.

Tilting of the piedmont is a necessary but not a sufficient condition for the formation of a pediment according to the conceptual model of this paper. In a relatively humid climate, bedrock is weathered more quickly than in an arid climate, resulting in a thick regolith and soil mantle on low-relief bedrock surfaces (e.g., the Appalachian Piedmont surface). Therefore, an additional requirement for pediment formation is that the rate of erosion must equal or exceed the rate of soil production on the piedmont. The rate of soil production is a function of climate (with lower rates in more arid climates), rock type, and soil cover. Therefore, given sufficient...
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Figure 2. Schematic diagram of the conceptual model. Early in the model, Basin and Range extension creates a semi-periodic series of basins and ranges. Range-bounding hillslopes and channels respond by backwearing and downwearing. The erosional response to uplift is accompanied by a flexural-isostatic rebound. If the value of $\alpha/\lambda$ is relatively large, isostatic rebound will be distributed across basins and range uniformly, resulting in little or no piedmont tilting (B). As a result, the piedmont will remain covered in alluvium. If, however, the value of $\alpha/\lambda$ is relatively small, the piedmont will tilt. Pediments will form on the tilted piedmont if the value of $P_d/\lambda x^2$ is relatively small. In that case (D), hillslope and channel erosion on the piedmont will keep pace with soil production, resulting in bare bedrock slopes despite the low relief (as in Fig. 1). If the value of $P_d/\lambda x^2$ is relatively large, a soil will form on the tilted piedmont slope. Also shown in (D) are parameters of the hillslope model (i.e., $s_p$, $\eta$, $\lambda$, and $h_0$) used to determine the critical value of $P_d/\lambda x^2$ required for bare slopes to form.

where $h$ is local elevation, $t$ is time, $U$ is rock uplift rate, $K$ is the coefficient of bedrock erodibility, $Q$ is discharge, $w$ is channel width, and $x$ is the along-channel distance (Whipple and Tucker, 1999). Scaling relationships between discharge, channel width, and drainage area can be used to further simplify (1) to

$$\frac{\partial h}{\partial t} = U - KA^m \frac{\partial h}{\partial x}$$  (2)

where $A$ is drainage area and $m$ (nominally equal to 0.5) is an exponent that combines the scaling relationships between discharge, channel width, and drainage area [note that the coefficient $K$ in (1) and (2) have different values and different units after (1) is transformed into (2)]. The more general form of the stream-power model includes a power-law relationship between erosion rate and channel slope (with an exponent often assumed to be close to unity, as assumed here) and a finite shear stress threshold for erosion (useful if a range of flood event sizes is prescribed). Here I use the basic form of the stream-power model exclusively because my purpose is to elucidate the nature of the feedbacks between bedrock channel incision, hillslope erosion, and flexural-isostatic rebound rather than to calibrate the stream-power model as precisely as possible for a particular study site.

Hillslopes in upland (soil over bedrock) landscapes are composed of a system of two interacting surfaces: the topographic surface $h(x,y)$, and the underlying weathering front, given by $h(x,y)$. The difference between these two surfaces is the soil or regolith thickness, $\eta(x,y)$. In this paper, the terms soil and regolith are used interchangeably to refer to the unconsolidated material above the bedrock consisting of material weathered in situ and material transported from upslope. The topographic and weathering-front surfaces are strongly coupled because the shape of the topography controls erosion and deposition, which, in turn, changes the values of $\eta(x,y)$ (Furbish and Fagherazzi, 2001). The values of $\eta(x,y)$, in turn, control bedrock weathering and/or soil production rates. The simplest system of equations that describes this feedback relationship between topography, soil thickness, and the rate of increase of soil thickness is given by:

$$\frac{\partial \eta}{\partial t} = \frac{P_r}{\rho_s \cos \theta} e^{-t \cos \theta h_0} + D \nu^2 h,$$  (3)

$$\frac{\partial b}{\partial t} = -\frac{P_r}{\rho_s \cos \theta} e^{-t \cos \theta h_0},$$  (4)

$$h = b + \eta.$$  (5)
where \( \rho_s \) is the bedrock density, \( \rho_c \) is the sediment density (nominally 1.3 times \( \rho_s \)), \( \rho_p \) is the maximum bedrock lowering rate on a flat surface, \( \Theta \) is the slope angle, \( D \) is the hillslope diffusivity, and \( \eta_\theta \) is a characteristic soil depth (Heimsath et al., 1997, 1999). Equation (3) states that the rate of change of soil thickness with time is the difference between a “source” term equal to the rate of soil production associated with the bedrock surface lowering and a “sink” term equal to the curvature of the topographic profile. The cos \( \Theta \) dependence in Equation (3) originates from the fact that soil production is an exponential function of soil thickness normal to the surface. The curvature-based erosion model in Equation (5) is the classic diffusion model of hillslope evolution, first proposed by Culling (1960, 1963). Equations (3–5) can be solved for the steady-state case in which soil thickness is independent of time:

\[
\eta = \frac{\eta_\theta \cos \Theta}{\left( \frac{\rho_s}{\rho_p} \right) \left( \frac{D \cos \Theta - \sqrt{\eta}}{\rho_m - \rho_c} \right)}.
\]

Note that steady state in this context does not mean that the topography is in steady state, but, rather, that the soil thickness is steady through time as the landscape is denuded (i.e., a “soil-thickness steady-state” condition). Equation (6) illustrates that, if the curvature is negative and greater than a certain threshold value, the surface will be bare of soil because the argument of the natural logarithm will approach one. The analysis of in situ cosmogenic isotope abundances indicates that the value of \( \eta_\theta \) (the soil thickness at which bedrock lowering falls to 1/e of its maximum value) is ~0.5 m for several well-studied sites around the world (e.g., Heimsath et al., 1997, 1999).

The transition between hillslopes and channels in the model occurs where the product of the slope and the square root of contributing area is greater than a threshold value given by the inverse of a channelization threshold \( X \) that has units of one over length (so that \( X \) has the same units as drainage density), i.e., \( SA^{1/2} \geq X^{-1} \). This approach is consistent with the slope and area controls on channel head location observed in natural drainage basins (e.g., Montgomery and Dietrich, 1988). The value of \( X \) is the square root of the average “support area” required to create a channel head on a particular landscape, per unit slope gradient. The value of \( X \) can be closely approximated by the drainage density of the basin divided by the average slope immediately downslope from channel heads. On the piedmont, hillslope and channel erosion are applied to all locations where the slope is greater than a threshold value given by \( S_{\text{min}} \). Below that value, the piedmont is assumed to store alluvium locally and hence does not erode.

The flexural-isostatic response to erosion is included in the model by solving for the deflection of a lithosphere with uniform elastic thickness subject to vertical unloading:

\[
DV^2w + (\rho_m - \rho_r)gw = q(x, y),
\]

where \( w \) is the deflection, \( D \) is the flexural rigidity, \( \rho_m \) is the density of the crust, \( \rho_c \) is the density of the mantle, \( g \) is the acceleration due to gravity, and \( q(x, y) \) is the weight of the rock removed by erosion (Watts, 2001). Elastic thicknesses in the Basin and Range vary from ~3 to 15 km (Lowry et al., 2000). Given a prescribed elastic thickness value, the flexural rigidity, \( D \), can be computed using the relationship:

\[
D = \frac{E}{12(1 - \nu^2)},
\]

where \( E = 70 \) GPa and \( \nu = 0.25 \) (typical values for continental lithosphere). Alternatively, the flexural-isostatic calculation can be completely described with two parameters: the relative density contrast between the crust and mantle, defined as \( (\rho_m - \rho_c)/\rho_m \) (nominally 0.2, e.g., \( \rho_c = 2750 \) kg/m\(^3\) and \( \rho_m = 3300 \) kg/m\(^3\)), and the flexural parameter, \( \alpha \), defined as:

\[
\alpha = \left( \frac{4D}{(\rho_m - \rho_c)/g} \right)^{1/4}.
\]

The flexural parameter is a length scale proportional to the natural flexural wavelength of the lithosphere (Turcotte and Schubert, 2002). Values of \( \alpha \) are approximately two to three times larger than those of elastic thickness (Pelletier, 2008). As such, appropriate values of \( \alpha \) in the Basin and Range vary from ~6 to more than 20 km (with lower values in areas of greater fault density or in close proximity to faults). Equation (8) is solved in the model using the Fourier transform technique (Press et al., 1992; Pelletier, 2008) validated by comparison of the model prediction with analytic solutions for deflection beneath a line load (Turcotte and Schubert, 2002).

The reference case of the model is a vertically uplifting mountain block 12.8 km × 12.8 km in extent. The model domain is 12.8 km in the model, i.e., comparable in relief to the natural flexural wavelength of the lithosphere. As such, the range value varies inversely with this parameter. As such, an appropriate value of \( K \) can be chosen so that the time scale of denudation down to base level is within the range of 30 to 50 Ma, thereby creating a mature mountain range at the end of 30 Ma of uplift and erosion that still stands high above base level. This constraint provides a reference value for \( K \) equal to 0.0002 km\(^{-1}\). Drainage densities in arid regions are sensitive to vegetation cover and hence precipitation. Melton (1957) found that drainage densities in southern Arizona are typically within the range of \( X = 0.01–0.1 \) m\(^{-1}\), with higher values in areas of greater aridity. Conceptually, the value of \( X \) can be thought of as the inverse of the average lateral distance between a divide and a channel head for a hill-slope with a gradient of 1 m/m (i.e., 45°). As such, the range \( X = 0.01–0.1 \) m\(^{-1}\) corresponds to channels with a spacing of between 10 and 100 m on steep slopes. For the reference model, I choose \( X \) to be at the high end of this range, i.e., 0.1 m\(^{-1}\). Pelletier and Rasmussen (2009) quantified the relationship between \( P_r \), mean annual temperature, and mean annual precipitation for granitic rocks and found \( P_r \) to be in the range of 0.01–0.1 m/ka in arid climates. In the reference case of the model, I assume \( P_r = 0.01 \) m/ka, but this parameter is also varied to determine the model sensitivity to variations in this parameter. Cosmogenic isotope studies constrain \( \eta_\theta \) to be ~0.5 m (Heimsath et al., 1997, 1999). Hillslope diffusivities in arid terrain range from \( D = 1–10 \) m/ka depending on hillslope sediment texture, vegetation cover, and other factors. I chose \( D = 10 \) m/ka for the reference case. Pediments form over a range of slopes from nearly flat to ~10° (Strudley and
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Murray, 2007). I choose the minimum slope for pediment formation, $S_{\text{min}}$, to be 0.02, but this value can be expected to vary with alluvial texture and drainage basin size. The value of $\alpha$ was chosen to be 12 km for the reference case but was varied between 9 and 24 km, with lower values expected in areas of the Basin and Range with higher tectonic activity. The conceptual model allows for the following criterion for pediment formation:

$$\frac{R^2}{\lambda} > S_{\text{min}}.$$  

Equation (13) provides an estimate for the largest value of $\alpha$ that will allow pediments to form in a Basin and Range system with prescribed values of $R$, $\alpha$, $\lambda$, and $S_{\text{min}}$. Figure 3 plots $\alpha_c$ as a function of $\lambda$ and $S_{\text{min}}$ assuming $R = 10$ km. Pediments can form for a wider range of values of $\alpha$ as $\lambda$ increases (i.e., ranges are more widely spaced) and $S_{\text{min}}$ decreases (e.g., sediments eroding from the mountains can be transported at lower slopes due to finer sediment textures and/or a wetter climate). Pediments can form in this simplified model, if the value of $\alpha$ is below that of $\alpha_c$. They will generally not form if $\alpha$ is greater than $\alpha_c$ because insufficient tilting takes place. Solutions of simplified models that help narrow the range of tectonic and climatic conditions under which pediments can form are further explored using approximate analytic solutions of simplified models that help narrow the range of tectonic and climatic conditions under which pediments can form.

Before describing the results of the fully coupled numerical model, the conceptual model can be further explored using approximate analytic solutions of simplified models that help narrow the range of tectonic and climatic conditions under which pediments can form. First, I consider the role of the flexural-isostatic response to erosion. Pediments can form, if flexural-isostatic tilting causes a portion of the piedmont to be steepened beyond $S_{\text{min}}$, the slope below which alluvial deposition is assumed to occur. The piedmont slope that results from tilting is equal to the ratio of the total rock uplift, $R$, divided by half of the Basin and Range wavelength, $\lambda/2$. Mathematically, this gives the following criterion for pediment formation:

$$R^2 > \frac{\lambda}{2} C > S_{\text{min}}.$$  

The factor $C$ in (10) is the compensation ratio, defined as the ratio of subsidence and/or uplift that results from lithospheric loading and unloading (Turcotte and Schubert, 2002). Mathematically, $C$ is defined as

$$C = \frac{1}{1 + \left[\frac{\pi \alpha}{4 \lambda} \right] T}.$$  

If the value of $C$ is equal to one, erosion in the range will result in uplift of the range only. As the value of $C$ decreases (by an increase in the value of $\alpha/\lambda$), isostatic rebound becomes more broadly distributed. If the value of $C$ is close to zero, isostatic uplift occurs with equal magnitude in basins and ranges, and little or no tilting occurs. The total rock uplift $R$ of the range during the interval from active uplift to erosion back down to base level is given by

$$R = \frac{\rho_m - \rho_e}{\rho_m - \rho_c} UT_c,$$  

where $U$ is the rate of active uplift, and $T_c$ is the duration of active uplift. For the model experiments of this paper, I assume $UT_c = 2$ km and $(\rho_m - \rho_c)/\rho_c = 0.2$, $\alpha = 12$ km, $U = 1$ m/ka, $T_c = 2$ Ma, $K = 0.0002$ ka$^{-1}$, $X = 0.1$ m$^{-1}$, $P_m = 0.01$ m/ka, $\eta_0 = 0.5$ m, $D = 10$ m$^2$/ka, $S_{\text{min}} = 0.02$, $\rho_m/\rho_c = 1.3$, and $m = 0.5$.

**DESCRIPTION AND RESULTS OF SIMPLIFIED MODELS**

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where $U$ is the rate of active uplift, and $T_c$ is the duration of active uplift. For the model experiments of this paper, I assume $UT_c = 2$ km and $(\rho_m - \rho_c)/\rho_c = 0.2$, so $R = 10$ km. The factor $\rho_m/(\rho_m - \rho_c)$ in (12) results from the fact that mountain ranges in isostatic equilibrium have buoyant crustal roots that are approximately five times thicker than the height of the overlying ranges. In order to erode mountains to base level, therefore, it is necessary to remove a thickness of rock equal to approximately five times as much as the height of the range. It should be noted that the amount of rock uplift in (12) is a maximum value. If pediments form, they will usually form prior to the complete denudation of the range.

Substituting (11) and (12) into (10) yields the following formula for the critical flexural parameter for pediment formation:

$$\alpha_c = \frac{\lambda}{2} \left( \frac{4\alpha}{S_{\text{min}}} - 4 \right)^{1/4},$$  

where

$$a = \frac{2UT_c\rho_c}{\lambda (\rho_m - \rho_e)}.$$  

Figure 3 plots $\alpha_c$ as a function of $\lambda$ and $S_{\text{min}}$ assuming $R = 10$ km. Pediments can form for a wider range of values of $\alpha$ as $\lambda$ increases (i.e., ranges are more widely spaced) and $S_{\text{min}}$ decreases (e.g., sediments eroding from the mountains can be transported at lower slopes due to finer sediment textures and/or a wetter climate). Pediments can form in this simplified model, if the value of $\alpha$ is below that of $\alpha_c$. They will generally not form if $\alpha$ is greater than $\alpha_c$ because insufficient tilting takes place to exhume the bedrock surface above the alluvial apron.

In addition to sufficient tilting, the climate must also be sufficiently arid and/or the lithology must be sufficiently weathering-resistant to prevent a thick mantle of soil and regolith from forming on the pediment surface. Pediment surfaces exhumed above the $S_{\text{min}}$ threshold will erode by bedrock channel and hillslope processes. Figure 2D illustrates the geometry of pediment hillslopes schematically. Soil production and erosion on hillslopes will occur via (3–5). The conditions required for bare slopes to form can be estimated by assuming
a soil-thickness steady-state condition. In this condition, the soil thickness is given by (6). The curvature of hillslopes on the pediment depends on the hillslope relief, \( h \), and the spacing between channels, \( \lambda_c \), according to
\[
\nabla^2 h = \frac{4h}{\lambda_c^2} \tag{15}
\]

Channel heads are located where the product of the local slope and the square root of contributing area is greater than a threshold value given by \( X \). This gives the following relationship for \( \lambda_c \):
\[
\lambda_c = \left( \frac{S}{X} \right)^{1/2} \tag{16}
\]

where \( S \) is the piedmont slope. Equation (16) assumes that \( \lambda_c \) is equal to the square root of the contributing area. Depending on the shape of the zeroth-order drainage basin above each piedmont channel head, \( \lambda_c \) may differ from \( A^{1/2} \) by a constant factor close to one [e.g., \( \pi(2)^{1/2} \) for the case of a semicircular zeroth-order basin]. By neglecting this factor, the results we obtain are only approximate. Substituting (15) and (16) into (6) gives
\[
\eta = \eta_0 \ln \left( \frac{P_s}{\rho_p} \frac{P_c}{4Dh_0S_c^{1/2}X^{2}} \right). \tag{17}
\]

The \( \cos \theta \) factors in (6) have been neglected in (17) because the piedmont slopes are sufficiently small that the approximation \( \cos \theta = 1 \) holds to a high degree of accuracy. Setting \( \eta \) equal to zero gives the following critical condition for the value of \( P_c/\lambda_cDX^2 \) required for bare slopes on the piedment:
\[
\frac{P_c}{\lambda_cDX^2} = \frac{\rho_p}{\rho_s} 4h_0S_c^{1/2}. \tag{18}
\]

For example, given the parameters of the reference model (i.e., \( P_s = 0.01 \) m/ka, \( D = 10 \) m/ka, \( X = 0.1 \) m\(^3\)), pediments will form (given sufficient tilting), if the right-hand side of (18) is greater than 0.1. This could occur, for example, with \( h_0 = 10 \) m and \( S_c = 0.05 \), assuming \( \rho_p/\rho_s \) is approximately one.

The simplified model results of this section highlight the importance of a sufficiently high degree of flexural-isostatic tilting and a sufficiently low rate of soil and regolith production in order for pediments to form. In the next section, I consider the results of the fully coupled model, which contains no approximations and illustrates how channel, hillslope, and flexural-isostatic processes interact to form pediments when certain tectonic, climatic, and lithologic conditions are met.

**MODEL RESULTS**

Figure 4 illustrates the results of the reference case numerical model. Figure 4A presents a color map of the topography following 2 Ma of active uplift and an additional 8 Ma of erosion and flexural-isostatic response to erosion following the cessation of active uplift. In this early phase of the model, a high plateau is still present, and bedrock channels have not had sufficient time to propagate back into the central portion of the range. Nevertheless, erosion has triggered a modest degree of isostatic rebound, as illustrated by the “halo” of dark-red colors surrounding the range. The piedmont slopes remain sufficiently small that no portion of the piedmont has tilted above the \( S_{\text{min}} \) threshold for subaerial exposure. As such, the piedmont is entirely covered with alluvium at this stage.

By \( t = 20 \) Ma (Figs. 4B and 4D), headward-eroding bedrock channels have penetrated into the headwaters of the range, and sufficient isostatic rebound has taken place to tilt the piedmont above the \( S_{\text{min}} \) threshold required for subaerial exposure of the initially suballuvial bedrock bench. Drainage networks have formed on the piedmont with second-order channels draining parallel to the direction of flexural-isostatic tilting. Periodically spaced, first-order channels have formed perpendicular to those second-order channels. The color map of soil thickness at \( t = 20 \) Ma illustrates that soil cover thickens with increasing distance downslope on the piedmont from zero or nearly zero on interfluves in the proximal end of the piedmont to greater than 3 m on the distal end of the piedmont. At later stages of the model (i.e., \( t = 30 \) Ma, Figs. 4C and 4E), erosion in the mountain range and the resulting flexural-isostatic rebound continue, but at a slower rate. As a result, soils are thicker everywhere on the piedmont at this time. It should be noted that white areas in the mountain range do not have thick soils despite the fact that they are mapped as white in the soil thickness maps of Figures 4D and 4E. This is because areas above 10% slope in the model were mapped as white regardless of the thickness of soil cover so that areas of thin soil (i.e., black areas) could be directly interpreted as pediments in these soil thickness maps.

For the purposes of this paper, it is necessary to choose a critical soil and regolith thickness below which the surface is considered a pediment and above which it is not. Here I choose \( \eta = 0.1 \) m as a threshold value, i.e., a surface with essentially no soil. This choice is consistent with the pediments in southern Arizona, many of which have essentially no soil cover (Tuan, 1959). Based on the literature, however, pediments with a range of soil thicknesses have been considered pediments, and hence one could choose \( \eta = 1 \) m, for example, as an alternative since there is no widely accepted threshold value. Clearly, the higher one sets the maximum soil thickness before a piedment surface is no longer considered a pediment, the wider the range of tectonic and climatic conditions under which pediments can form. For the purposes of interpreting the model results, however, the important thing is to use a consistent threshold value for soil and regolith cover across the range of model scenarios so that the relative sizes of the pediments formed in each model can be compared and the controls on pediment area properly identified.

Figure 5 is a plot of the pediment area as a function of time in the reference model and several alternative models, illustrating the sensitivity of pediment area to variations in individual parameter values and time. In the reference model, pediment area increases with time until a maximum area is reached at \( t = 17 \) Ma and then declines back down toward zero. The time scale of the waxing and waning of pediment area in the model is primarily a function of the bedrock erodibility, \( K \), which is the primary control on the time scale of mountain range denudation. The sensitivity of pediment area to individual model parameters is illustrated in Figures 5 and 6. First, I increased the value of \( \alpha \) from 12 to 18 km. Tilting in the model with \( \alpha = 18 \) km was insufficient to steepen any portion of the piedmont above \( S_{\text{min}} \). As a result, no pediment was formed (Fig. 6B). This extreme sensitivity of pediment area to \( \alpha \) can be understood as a consequence of the fact that the flexural-isostatic response of the lithosphere is very sensitive to \( \alpha \) (i.e., \( C \) varies inversely as the fourth power of \( \alpha/\lambda \)) coupled with threshold nature of pediment formation in the model. Second, increasing the value of \( P_s \) from 0.01 m/ka (the reference case) to 0.03 m/ka thickens soils on the piedmont and thereby decreases pediment area (Figs. 5 and 6C). Third, decreasing the channelization threshold from \( X = 0.1\)–0.05 m\(^3\) results in smaller pediments. In the mountain range, a lower channelization threshold slows erosion because the ratio of hillslopes to channels increases with decreasing \( X \), and hillslopes erode more slowly than channels. As a result, isostatic rebound is also reduced. On the piedmont, lower drainage densities result in thicker soils, thereby resulting in a further reduction in pediment area in the \( X = 0.05 \) m\(^3\) model relative to the reference case. Finally, I also considered the effect of reducing the active uplift rate \( U \) by a factor of 5 but lengthening the duration of active uplift \( T_u \) by the same proportion so that the total active rock uplift remained the same as in the reference case. A smaller value of \( U \) leads to a modest
How do pediments form?

Pediments in the model can form by exhuming the suballuvial bench downslope from the range-bounding fault and/or by retreat of the mountain front upslope from the range-bounding fault. Examples of both mechanisms are found among the pediments of the Mojave and Sonoran Deserts. In the model, the relative importance of each mechanism depends sensitively on the value of \( P_0 \). On the hillslopes of the mountain range, soil cover is negligible due to the steep slopes and relatively low values of \( P_0 \) used in the model. As such, slopes erode normal to the hillslope with a rate nearly equal to the maximum rate of bedrock weathering, \( P_0 \). The mountain front will therefore retreat laterally with a rate of approximately \( P_0 \sin \theta \), where \( \theta \) is the hillslope angle. Figure 8 illustrates the model topography at \( t = 20 \) Ma for four different values of \( P_0 \), keeping all other parameters of the model equal to those of the reference case. If \( P_0 = 0.03 \) m/ka (Fig. 8A), and the angle of the hillslopes along the mountain front is assumed to be 45°, the mountain front wears back, extending the pediment, by ~423 m during the 20 Ma duration of the model illustrated in Figure 8. If \( P_0 = 0.05 \) m/ka (Fig. 8B), the mountain front is expected to wear back by 705 m in the same time period. The grayscale map of Figure 8B, however, indicates that more than 1 km of mountain front retreat has taken place in the \( P_0 = 0.05 \) m/ka case. Similarly, if the value of \( P_0 \) is raised to 0.07 m/ka, at least 2 km of mountain

Figure 4. Color maps of output of the numerical model for the reference case. (A) Color map of topography at \( t = 10 \) Ma (model starts at \( t = 0 \) Ma and ends at \( t = 30 \) Ma). Flexural-isostatic rebound is just beginning, as indicated by the “halo” of dark red surrounding the mountain block. (B–E) Color maps of the topography (B and C) and soil thickness (D and E) at \( t = 20 \) Ma and 30 Ma, respectively, illustrate the maximum and waning stages of pediment formation. In (B) and (D), the pediment area (shown in black, indicating less than 10 cm soil cover) is ~25% of the model domain. At later times where the upland topography is more subdued, the pediment has retreated toward the mountain as soil thicknesses have increased.

This result can be interpreted as a consequence of a reduction in mountain front hillslope gradients with lower values of \( U \). Mountain front steepness promotes greater slope backwearing relative to downwearing, thus lengthening the pediment surface by slope retreat.

The trends among pediment areas \( \alpha \), \( P_0 \), and \( X \) discussed above are also documented in Figure 7, which illustrates the relationship between pediment area \( A_p \) (expressed as a fraction of the model domain), \( \alpha \), and \( P_0 \) for two representative values of channelization threshold, \( X \). Pediment area decreases steadily with increasing values of \( \alpha \) and \( P_0 \), in the former case because not enough piedmont tilting takes place and in the latter case because there is too much soil cover on the piedmont. It should be emphasized that the specific thresholds for pediment formation documented in Figure 7 are specific to \( \lambda = 25.6 \) km and other model parameters. As such, while it is reasonable to expect that the dependence of \( A_p \) on \( \alpha \) and \( P_0 \) documented in Figure 7 will hold qualitatively for other parameter values, the specific thresholds for pediment formation will differ from those of Figure 7, if any of the other model parameter values deviate from those of the reference case.
Figure 5. Plot of time series of pediment area $A_p$ expressed as a fraction of the area of the model domain, for the reference case and three alternative models with lower $U$, lower $X$, and higher $P_0$. In each case, pediment area increases as incipient canyons erode headward thereby triggering flexural-isostatic rebound. After a maximum pediment area is achieved (ranging from $t = 15$–25 Ma), the rate of flexural-isostatic rebound decreases and soils accumulate on the piedmont. The model with $\alpha = 12$ km is not shown because the plot is indistinguishable from $A_p = 0$ (i.e., the x axis).

front retreat takes place in the same time interval. The reason for this apparent nonlinear increase in slope retreat rate with $P_0$ can be traced to the fact that once the mountain front becomes embayed by bedrock channels, slopes retreat back into the mountain range and laterally away from the incised bedrock canyons. The effect of an embayed mountain front on the effective rate of mountain front retreat is illustrated schematically in Figure 8D, where the retreat of a linear mountain front is compared to that of a sinuous mountain front in map view. In an embayed mountain front, slope retreat from two adjacent embayments can sum together to retreat the slope back at a rate that exceeds $P_0 \sin \theta$. In this figure the length of the arrow between the dashed lines, which represents the effective retreat rate of the embayed mountain front, is longer than the length of all other arrows in the figure, which represent the rate $P_0 \sin \theta$. This effect can also be understood in terms of an increase in the overall length of the mountain front over time. $P_0 \sin \theta$ is the rate of slope retreat per unit mountain front length. As the mountain front lengths and develops embayments, the effective rate of mountain front retreat must also increase. The results illustrated in Figure 8 are broadly consistent with empirical studies of embayment formation in arid regions. Parsons and Abrahams (1984), for example, documented the importance of slope retreat in contributing to the formation of mountain front embayments in the Mojave and Sonoran Deserts.

The assumption of uniform vertical uplift is clearly an idealization. In order to determine the effect of tectonic tilting on the distribution of pediments around a mountain range, I performed a numerical experiment in which the active uplift was prescribed to have a uniform “background” value of 0.5 m/ka and an additional asymmetric tilt component with a maximum value of 0.5 m/ka, thereby yielding a maximum rate of $U = 1$ m/ka (i.e., equal to that of the reference case). As in the earlier experiment performed with $U = 0.2$ m/ka, the duration of uplift was lengthened so that the total uplift integrated over the mountain range was equal to that of the reference case. In the meta-morphic core complexes of southern Arizona, asymmetric tectonic tilting has played a significant role in the development of drainage architecture (Pelletier et al., 2009), and its role in pediment formation is suggested by asymmetric pediment development on the margins of many metamorphic core complexes of southern Arizona. Figure 9 illustrates the topography and soil thickness predicted by this asymmetric tilt block model. Pediments are best developed on the north, west, and east sides of the mountain range, i.e., the directions opposite to the direction of tilting. This result can be understood as a consequence of the fact that the center of mass of the eroded material has shifted northward in this case, thereby shifting the maximum isostatic rebound and center of tilting northward as well.

In the model, I assume the bedrock surface exhumed from beneath the suballuvial bench is flat. This surface need not be flat, however, and hence it is reasonable to ask whether exhumation of a rough bedrock surface from beneath some initial cover of alluvium will also form a pediment. Model experiments with an initially rough bedrock surface indicate that the answer is yes, a pediment still forms. The reason a pediment still forms is that, even if the bedrock surface is initially rough, the large-scale relief of the surface in either case (initially flat or initially rough) is controlled by the magnitude of flexural-isostatic rebound, not by the initial relief of the buried surface. In the case of an initially rough surface, only the top-most portions of the surface are exhumed at first as flexural-isostatic tilting begins, leaving lower portions still buried under alluvium. This situation is not significantly different from the case of an initially flat bedrock surface, although in the case of a rough surface the pediment may be spatially discontinuous. In the case of an initially flat bedrock surface, flexural-isostatic tilting triggers bedrock channel incision, increasing relief and/or slopes at the hillslope scale. A pediment forms, if relief reduction (by weathering and/or slope retreat) occurs sufficiently fast to balance relief production by bedrock channel incision. In the case of an initially rough bedrock surface, relief already exists at the hillslope scale (and hence may not be increased by bedrock channel incision), but the large-scale relief is the same in both cases because it is controlled by the magnitude of flexural isostatic tilting above the level set by $S_{max}$, the gradient at which alluvium is deposited, not the initial relief of the buried surface. It should also be noted that the ability of weathering to reduce the relief of slowly tilted landscapes (whether the initially buried surface is flat or rough) does not require the presence of regolith. Even in the absence of regolith cover, relief reduction occurs because weathering takes place normal to the surface, thereby including a $1/\cos \theta$ component that causes high-relief areas to be eroded more rapidly than low-relief areas.

**COMPARISON TO PEDIMENTS IN SOUTHERN ARIZONA**

Southern Arizona is home to many of the classic pediments of the southwestern United States. Descriptive studies by Paige (1912), Bryan (1922), Gilluly (1937), and Tuan (1959), among others, have made the pediments of this region type localities. In this section, I use geographic information system (GIS) data sets and field observations to relate the predictions of the numerical model to observed trends in the occurrence and size of pediments in this region.

In order to map pediments within a GIS framework, it is necessary to develop geographically co-registered data sets for topographic slope and the presence and/or absence of crystalline bedrock. Pediments can then be mapped by identifying all pixels with slopes less than a certain threshold value (nominally 10%) that are also composed of crystalline bedrock. To obtain a slope map for southern Arizona, I began with a 90 m/pixel digital elevation model (DEM) of southern Arizona obtained from the U.S. Geological Survey (USGS). U.S. Geological Survey DEMs have well-known slope artifacts associated with contour lines. To minimize this problem, I averaged the slope map produced from the DEM using a $5 \times 5$ (or $450 \times 450$ m) moving window.
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Average. Other smoothing algorithms could also be used, but the 5 x 5 pixel moving average worked adequately at minimizing contour artifacts while also minimizing the artificial steepening of piedmonts close to mountain fronts that is unavoidably introduced by smoothing. A “mask” grid of the presence and/or absence of crystalline bedrock was obtained by projecting the Arizona digital geologic map (Hirschberg and Pitts, 2000) to the same projection and resolution as the DEM. The Arizona digital geologic map contains data on the lithology and age of all bedrock and sediment in the state resolvable at 1:500,000 scale. For the purposes of this analysis, I mapped crystalline bedrock as all pre-Cenozoic map units that are not sedimentary rock. In southern Arizona, nearly all basalt, sediment, and sedimentary rocks are Cenozoic in age. As such, the pre-Cenozoic criterion works well for distinguishing crystalline rocks (excluding basalt) from sedimentary rocks in this region. Planation surfaces can form in sedimentary and basaltic rocks, but these are usually associated with the differential erosion of strata or basalt flows, and hence are not included within the definition of pediments adopted in this paper. Some granites in southern Arizona are Cenozoic in age, so this map accounts for most but not all of the granite in this region. Figure 10A illustrates a shaded relief image of the DEM. The pediment map (Fig. 10B) obtained by identifying all crystalline bedrock areas with a slope less than or equal to 10% is illustrated in Figure 10B. Pediments of one size or another occur adjacent to nearly all mountain ranges in southern Arizona. In this paper I chose to highlight two specific pediment regions for discussion: the Catalina and Oracle pediments on the west and north sides of the Santa Catalina Mountains, respectively, and the Sierrita pediment on the north side of the Sierrita Mountains (locations shown in Fig. 10A).
The pediments of the Santa Catalina Mountains are best developed on the west and north sides of the range (Fig. 11). These pediment surfaces are composed of the same granite that makes up the steeper portions of the range. The pediment surface is dissected by bedrock channels that are locally incised by as much as tens of meters. These channels are also clearly influenced by the jointed structure of the Catalina granite (Bezy, 1998; Pelletier et al., 2009; Fig. 1B). Figure 1B indicates that the pediment surface is characterized by a high drainage density, i.e., despite the gentle slope of the surface (<10%), channels are separated by ~10 m, implying values of X in the range of 0.1–1 m⁻¹. The high drainage density on this surface may be the result of a combination of the lack of regolith and the jointed nature of the bedrock. On the east side of the range, late Oligocene–early Miocene sedimentary rocks have been exhumed along the mountain front following a cycle of syntectonic sediment deposition, burial, and exhumation. Most of the piedmont on the east side of the range is a planation surface, and hence the pediment is restricted to a small area on this side of the range because of this lithologic contrast.

The Santa Catalina Mountains are primarily composed of granite and to a lesser extent by mylonitic gneiss on the south side of the range. The Santa Catalina Mountains are bounded on the south side by the low-angle Catalina detachment fault (Fig. 11A) and on the west side by the high-angle Pirate fault (Figs. 1B and 11A). Offset along these faults occurred in two separate intervals of deformation and uplift. In the initial phase, late Oligocene–early Miocene offset occurred primarily along the Catalina detachment fault. Extension along an ~240° azimuth was accompanied by tectonic tilting of an extension-parallel topographic ramp and by antiform arching along a direction approximately orthogonal to extension (Dickinson, 1991). In the second, post–mid-Miocene tectonic event, faulting occurred along the high-angle Pirate fault in a manner similar to the block faulting associated with classic Neogene Basin and Range extension (Davis et al., 2004; Wagner and Johnson, 2006). The initial phase of extension was accompanied by tilting to the southwest, resulting in relatively steep western, northern, and eastern edges of the range and a relatively gently dipping south and southwestern flank characterized by larger drainages. The model results illustrated in Figure 9 provide a basis for interpreting the asymmetry of the pediments flanking the Santa Catalina Mountains in terms of the south and southwesterly tilting of the range during the initial phase of extension. In the model of Figure 9, south-directed tilting leads to the development of steep mountain fronts and well-developed pediments on
How do pediments form?

The Sierrita Mountains were formed during the same late Oligocene–early Miocene extension that uplifted the Santa Catalina Mountains (Stavast et al., 2008). They did not take part in late Miocene high-angle faulting and uplift, however, and hence they have lower relief. The Sierrita Mountains pediment differs from those of the Santa Catalina Mountains in that it is characterized by relief on the order of meters rather than tens of meters and because it has a discontinuous cover of alluvium that thickens downslope into an alluvial apron (Bengert, 1981). As such, the boundary between the pediment and the alluvial apron is gradational in this case (Fig. 11). Tuan (1959) argued that the gradational nature of this contact suggests that the Sierrita Mountain pediment is an exhumed suballuvial bedrock bench. More generally, Tuan concluded, “careful observation of the pediments of southeastern Arizona reveals that most of them bear evidence of exhumation” (Tuan, 1959, p. 124). The Sierrita pediment is more similar to the results of the model illustrated in Figure 4 than is the Catalina pediment. In the model, pediments are formed on the proximal end of the piedmont, giving way to thicker soils and ultimately an alluvial apron with increasing distance downslope. The Catalina pediment, in contrast, is bounded abruptly on its eastern side by the Pirate fault. The Sierrita pediment is best expressed on its northern side (Fig. 10B). This fact can be understood, in part, as a consequence of the fact that pediments require a sufficiently low value of $\alpha/\lambda$ to generate sufficient tilting to become erosional surfaces. If a range is separated from its neighboring ranges by varying distances, the model of this paper predicts that pediments will be most well developed on the side that is separated from its neighboring ranges by the largest distance (hence minimizing the value of $\alpha/\lambda$, assuming uniform $\alpha$). In the case of the Sierrita Mountains, the northern side of the range has the greatest distance from surrounding ranges, hence the pediment is expected to be most well developed on that side of the range.

The flexural-isostatic modeling of this paper can be tied more explicitly to the landscapes of southern Arizona by calculating the flexural-isostatic response to erosion in southern Arizona. The erosion rates of different ranges are not precisely known, of course. Nevertheless, it is possible to assume a uniform rate of erosion in all of the steep (>10%) portions of the region and then to model the relative rates of flexural-isostatic rebound that would result from such erosion. Figure 12 illustrates color maps of the flexural-isostatic uplift in southern Arizona, expressed as a fraction of the total erosion, for different values of the flexural parameter $\alpha$. For $\alpha = 9$ km (Fig. 12B), isostatic rebound ratio approaches its maximum (Airy isostatic) value of ~0.8. For larger values of $\alpha$, the peak isostatic rebound (and hence degree of piedmont tilting) decreases as the flexural-isostatic response becomes more spatially distributed. For $\alpha = 17$ km (Fig. 12D), for example, the maximum compensation is approximately half of its value for $\alpha = 9$ km, and...
the isostatic response to erosion within some of the more narrow ranges in southern Arizona is distributed across one or more adjacent ranges and their intervening basins. Of the three maps presented in Figure 12, the map in Figure 12C (corresponding to $\alpha = 13$ km) produces a pattern of tilting most similar to the distribution of pediments in Figure 10B. This result lends support to the value of $\alpha$ (i.e., 12 km) chosen for the model reference case. The flexure maps in Figure 12 cannot be used directly to reproduce the pediment map of Figure 10B due to spatial variations in $\alpha$, mountain block erosion rates, etc. that are not well constrained using existing data.

**DISCUSSION**

Cooke (1970) performed a detailed statistical analysis of 53 pediments in the western Mojave Desert in an attempt to develop diagnostic relationships between the morphology of pediments and the mountain range drainage basins upslope from them. Cooke documented a poor correlation between pediment length and slope, a surprising result considering that pediments generally decrease in slope with increasing distance downstream (hence, one would expect a negative correlation between pediment length and slope). Cooke argued that longer pediments must have undergone greater tectonic tilting in order to explain this lack of correlation between pediment slope and length, a conclusion consistent with the model of this paper. Cooke (1970) also documented essentially no correlation between the area of the pediment and the area of the drainage basins upslope from them in the mountain range. If pediments expand primarily by backwearing of the mountain front, one would expect a negative correlation between pediment area and upstream drainage basin area. Cooke (1970) interpreted the lack of such a correlation to imply that pediments grow primarily by exhumation of the suballuvial bench, not by

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**Figure 11.** Pediment maps overlain on shaded relief images for (A) Santa Catalina Mountains and (B) Sierrita Mountains. (C and D) Aerial photographs of pediments on the (C) west and (D) north side of the Santa Catalina Mountains, illustrating the low-relief, dissected nature of these pediments. In contrast, the Sierrita Mountains pediments (E and F) have a thin, discontinuous veneer of regolith and alluvium with tors and inselbergs dotting the landscape (E).
How do pediments form?

Figure 12. (A) Thresholded slope map for southern Arizona, used as input to the flexural-isostatic model. (B–D) Color maps of isostatic rebound, expressed as a fraction of the total erosion in the ranges, for (B) \( \alpha = 9 \) km, (C) \( \alpha = 13 \) km, and (D) \( \alpha = 17 \) km. As the value of \( \alpha \) increases in the model, tilting is reduced (i.e., yellow and white colors, indicative of high magnitudes of isostatic rebound, are no longer present), and the isostatic rebound becomes more uniformly spatially distributed.

retreat of the mountain front. Cooke (1970) concluded: “the exhumation hypothesis for pediment development deserves close examination in the western Mojave Desert” (Cooke, 1970, p. 36). The role of flexural-isostatic rebound in pediment formation has not yet been proposed in the pediment literature. As such, the model of this paper fills an important gap in the exhumation hypothesis since it provides a widely applicable mechanism for piedmont tilting. At larger spatial scales, flexural-isostatic rebound has been recognized as an important process in “pediplanation.” For example, flexural isostasy has been invoked as a necessary component for the formation of the stepped landscape of southern Africa (Pugh, 1955).

The model of this paper provides a working hypothesis for the relative roles of tectonics, climate, and rock type in controlling the occurrence of pediments. The tectonic style of a region controls the values of \( \alpha \) and \( \lambda \). Regions that are more extensively faulted are characterized by lower values of \( \alpha \) and hence are more likely to form pediments under otherwise similar conditions. This is consistent with correlations between pediment occurrence and the presence of faults in the Mojave and Sonoran Deserts (Mammerickx, 1964; Cooke, 1970). The occurrence of faults that bound pediment surfaces cannot be explicitly accounted for in the modeling framework of this paper (which assumes a continuous elastic lithosphere). Nevertheless, the presence of pediment-bounding faults will decrease the value of \( \alpha \) locally, hence a prediction of the model is that pediments are more likely to form in areas of pervasive faulting. Pediment formation can also be influenced by the rate of active uplift, \( U \) (as illustrated in Fig. 5) and the angle of faulting. Higher uplift rates and steeper faults both encourage pediment formation because they promote greater channel and slope backwearing (which tends to lengthen the pediment) over downwearing (which does not).

Climate controls pediment formation principally through its control on the maximum rate of bedrock weathering, \( P_o \), Pelletier and Rasmussen (2009) quantified the relationship between \( P_o \), mean annual temperature, and mean annual precipitation for granitic rocks based on an analysis of cosmogenic-radiocarbon-derived erosion rates compiled from the literature. In figure 3 of that paper, they showed that \( P_o \) varies by several orders of magnitude between arid and humid conditions. For this reason, I emphasize the importance of climatic control on \( P_o \) as the principal role of climate on pediment formation even though climate also controls the values of \( S_{\text{min}} \), \( X \), and \( D \) in the model of this paper. Wetter climates increase surface runoff, most likely increasing \( D \) and decreasing \( S_{\text{min}} \) (higher discharges require lower threshold slopes to entrain sediment of a given texture). Wetter climates generally have lower drainage densities, \( X \), within arid to semiarid climates (e.g., Melton, 1957) because increasing vegetation density increases soil resistance to erosion. Given the sensitivity of \( X \) and especially \( P_o \) to climate, it is reasonable to expect that the climatic dependence of \( P_o \) and \( X \) (with greater aridity promoting pediment formation) will dominate over the tendency for decreases in \( D \) and increases in \( S_{\text{min}} \) to prevent pediment development in more arid climates.

Rock type also controls the values of \( P_o \), \( S_{\text{min}} \), \( X \), and \( D \). Rocks that are more resistant to weathering have lower values of \( P_o \), and hence are more likely to form pediments than less resistant rocks. Rock type also controls \( S_{\text{min}} \) via the texture of detritus that the rock weathers to (on the hillslope) and abrades to (in the channel) (Abrahams et al., 1985; Parsons and Abrahams, 1987). Rock types that minimize both \( P_o \) and \( S_{\text{min}} \) are ideal for pediment formation. The presence of granite, for example, facilitates pediment formation because it often weathers to small particles (grus) (Pye, 1986) and therefore results in lower values of \( S_{\text{min}} \) and possibly higher values of \( D \) (smaller particles are more readily transported).
Pediment formation can occur for a range of values of $P_r/\Delta X^2$. As noted above, if the maximum soil thickness for classifying a piedmont as a pediment is increased, e.g., from the 0.1 m used here to the 2–4 m used by Strudley and Murray (2007), the range of $P_r/\Delta X^2$ values conducive to pediment formation will also increase. More work is needed, however, to narrow the range of each parameter value, and their relationships with climate, rock type, and soil cover, to enable the model to be calibrated precisely for specific study areas. Pelletier and Rasmussen (2009), for example, calibrated the climatic dependence of $P_r$ for granite, but their work was based on a relatively small population of granites. Locally, $P_r$ values of granite will depend on biotite concentration, joint spacing, and other factors. Melton (1957) identified $X = 0.01–0.1$ m$^{-1}$ as an appropriate range of drainage densities in arid to semiarid climates, but the value of $X$ also likely depends on the soil thickness $\eta$ (i.e., soils devoid of regolith likely have greater runoff and therefore higher values of $X$ for otherwise similar conditions). If true, this raises the possibility of a positive feedback between the thickness of soil cover and the value of $X$ such that, once pediments form, drainage densities increase on the pediment surface, thereby increasing the ability of hilltops to transport regolith. Such a feedback would tend to preserve pediments once they form.

Evidence from cosmogenic radionuclide studies (Heimsath et al., 1997, 1999) indicates that the soil production function can take on the form of either an exponential or a “humped” function. Strudley et al. (2006) argued that if a humped production function is operable on a given piedmont, pediments are more likely to form because the humped production function lowers the value of the maximum soil production rate and, hence, results in thinner soils for otherwise similar conditions. While I agree with Strudley et al. (2006) that the humped production function can promote pediment development, I choose to focus the model of this paper on the exponential production function for two reasons. First, the mathematical form of the humped production function is not well constrained, so the magnitude of the difference in soil production rates between the exponential and humped functions for a given soil thickness is not well constrained. Second, I believe this difference is likely to be a second-order effect relative to the magnitude of natural variations in $P_r$. In the model of this paper, pediment surfaces can form as long as soil production occurs more slowly than soil erosion on piedmont surfaces. That condition can be expected to occur over a slightly narrower range of climatic, tectonic, and lithologic conditions using the exponential production function compared to a humped production function. Nevertheless, bare slopes will form in either case provided that the value of $P_r/(\Delta X^2)$ is sufficiently low.

The model of this paper has a number of limitations. Most importantly, the model does not treat erosion and deposition in a mass-conservative manner. Instead, the model simply assumes that alluvial fans will backfill on a surface that is below a certain critical slope given by $S_{\text{crit}}$. The slopes of alluvial fans are functions of time and the hinterland erosion rate in addition to sediment texture and climate (both of which are implicitly included in $S_{\text{crit}}$). As such, a more sophisticated approach is needed to honor mass conservation so that alluvial fan deposition can be more accurately represented in the model. Similarly, the model could be improved by incorporating faults via a model that forces the flexural rigidity to go to zero or nearly zero at fault locations.

CONCLUSIONS

Pediments have puzzled geoscientists for more than a century. Recent advances in numerical modeling, led by Strudley et al. (2006; Strudley and Murray, 2007) have rejuvenated interest in the debate over how pediments form and what factors control their morphology. The model of this paper couples bedrock channel erosion, soil production and erosion on hillslopes, and the flexural-isostatic response of the lithosphere to loading on hillslope) processes in addition to feedbacks between climate, tectonics, and erosion.

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REFERENCES CITED


How do pediments form?


