Influence of precipitation phase on the form of mountain ranges

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ABSTRACT

Observations of precipitation fall speeds and precipitation patterns suggest that precipitation phase (rain versus snow) is a significant control on the relationship between precipitation patterns and topography, due to the potential for increased downwind advection of snow relative to rain. A coupled model of orographic precipitation and surface erosion shows that for a range of climate variables, steady-state precipitation patterns vary from nearly uniform and maximizing over the highest topography, to highly spatially variable, closely coupled to topography and reaching a maximum on low slopes. Precipitation patterns are a first-order control on modeled range scale and ridge-valley scale relief, channel concavity, and the position of the drainage divide. An association between cool climates, spatially uniform precipitation, and efficient erosion of high topography is indicated. The importance of precipitation phase to the evolution of precipitation patterns and topography further demonstrates the fundamental importance of the coupled climate, erosion, and tectonic system in the evolution of mountain topography.

Keywords: surface processes, climate, numerical modeling, precipitation.

INTRODUCTION

Climate, in general, and precipitation, in particular, exert some control on nearly all erosional processes. Precipitation accumulated as river discharge is fundamental to landscape change and the geodynamic evolution of active orogens over millions of years (e.g., Whipple, 2004). Likewise, glacial erosion rates scale with ice discharge and thus with precipitation rate (Hallet, 1979; Oerlemans, 1984). Precipitation patterns in mountains are controlled by topography (e.g., Roe, 2005), and persistent spatial gradients in precipitation have been documented in the European Alps (Frei and Schär, 1998), Himalaya (Anders et al., 2006; Barros et al., 2006) and Olympic Mountains of Washington State (Anders et al., 2007). The influence of spatially persistent precipitation patterns on geologic evolution has received considerable attention (e.g., Montgomery et al., 2001; Reiners et al., 2003).

We investigate whether interaction between the first-order physics of orographic precipitation and large-scale erosion has the potential to fundamentally shape landscapes. We highlight the importance of precipitation phase (rain versus snow) as an influence on the interaction between precipitation patterns and topography. Changes in the dominant precipitation phase are explored with coupled numerical models of orographic precipitation and fluvial erosion, demonstrating that the resulting pattern of precipitation is a fundamental control on the modeled landscape. We emphasize that although the focus of this paper is to evaluate the potential for important interactions between precipitation phase, precipitation patterns, and topographic evolution, we also discuss specific measurements that could be made in natural settings that would address the extent to which the coevolution of precipitation patterns and topography is expressed in real landscapes.

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PRECIPITATION PHASE

Ascent upwind of topographic highs induces condensation of water vapor in saturated air. Initial nucleation of cloud droplets of ~10 µm in diameter is followed by growth of two orders of magnitude to reach the size of typical falling particles (e.g., Roe, 2005). When precipitation particles are large enough to have a downward-directed velocity, the terminal fall speed is dependent on precipitation phase: rain falls at 5–10 m/s, while snow falls an order of magnitude more slowly at 0.5–1 m/s. As precipitation particles are growing, they are advected downwind, typically at rates greater than the fall speed (~20 m/s). It is common for precipitation that falls on the surface as rain to have formed as snow and melted during descent. The fraction of time spent as frozen versus liquid precipitation strongly influences the advection distance. If it falls as rain for most of its descent, the precipitation generated by topographically induced lifting will be closely associated with topography. If precipitation falls as snow, it is advected much farther downwind and has a diffuse relationship with topography. We explore the importance of precipitation phase in determining the relationship between precipitation patterns and topography using the orographic precipitation model of Smith and Barstad (2004).

The linear orographic precipitation model computes precipitation rates under imposed climatic conditions: the air is assumed to be saturated through the troposphere and has an adiabatic vertical temperature profile; horizontal wind speed and direction are constant in space and time; and wind speed and direction, surface temperature, and the moist static stability of the atmosphere (i.e., the resistance to vertical displacement over topography) are imposed (Smith and Barstad, 2004). The linear flow of air over topography determines where and how rapidly air is ascending. In regions of ascent, water vapor condenses into precipitation particles. The processes of nucleation, growth, and fallout of precipitation particles are represented as a characteristic delay time (Smith and Evans, 2007).

The delay time can be tuned to simulate observed patterns of precipitation (Smith et al., 2005; Barstad and Smith, 2005; Smith and Evans, 2007). Case studies of storms in the Wasatch Range of Utah, the southern California

Coast Range, and in the European Alps have produced estimates of delay time (Barstad and Smith, 2005). The climatological (long-term) pattern of precipitation and isotopic depletion of precipitation have been used to constrain total delay times for the Oregon Coast Ranges and Cascades (Smith et al., 2005), the Andes (Smith and Evans, 2007), and the Olympic Mountains of Washington State (Anders et al., 2007). Together, these studies reveal a systematic relationship between the total delay time and the mean annual temperature (Fig. 1). These observations are consistent with the hypothesis that the increasing prevalence of snow in cool climates accounts for the increase in delay time needed to fit the observed precipitation patterns. This suggests that precipitation is more closely tied to local topography in warm climates than in cool climates. Unfortunately, there are no estimates based on climatological precipitation patterns from warmer (e.g., tropical) settings.

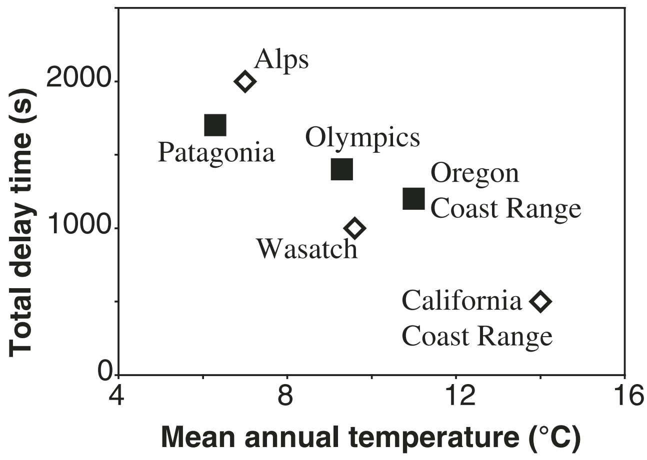

Figure 1. Inverse relationship between mean annual temperature and estimated total delay time; data were compiled from studies of individual events, shown as diamonds, and long-term studies, shown as squares.

COUPLED MODEL

We use a coupled model of landscape evolution and orographic precipitation to explore the impacts of changes in precipitation phase on the steady-state form of mountain ranges. This model combines the linear orographic precipitation model of Smith and Barstad (2004), with the landscape evolution model CASCADE (Braun and Sambridge, 1997). The fluvial incision model in CASCADE has been altered from Braun and Sambridge (1997) to represent fluvial incision as proportional to river discharge (i.e., basin-integrated precipitation) to the ½ power multiplied by slope. A limiting threshold slope of 30° is imposed. The model resolution is 1 km, which limits the area reaching the threshold slope to <2%. The domain is 256 km by 64 km with the long sides pinned at zero elevation. No transport is allowed through the short sides. The initially flat topography is seeded with small-amplitude noise (<10 m). Uplift is spatially uniform at 2 mm/yr except along the long sides of the domain. The surface process model is run with a 1 yr time step and the orographic precipitation model is run every 2000 model years. A relatively large minimum precipitation rate of 0.4 m/yr is imposed in order to keep the estimates of spatial variability in precipitation conservative. The model is integrated until it reaches steady state, which occurred by 10 m.y. in all cases.

The use of a simple fluvial-incision model allows us to isolate the effects of spatially variable precipitation without additional factors such as variability in discharge or effects of sediment in the river channel. These factors are likely important in real systems, for some time and length scales, but are neglected here for simplicity. We emphasize that these numerical experiments are not meant as simulations of particular ranges, but as exploratory steps toward understanding the coupled climate-erosion system. Limited testing confirms that results remain qualitatively similar when other fluvial incision laws are used (shear stress and total stream power) (Anders, 2005).

To investigate the role of precipitation phase in the evolution of the coupled system, we vary the delay time in the precipitation model and compare the resulting steady-state landscapes and precipitation patterns. The orographic precipitation model is sensitive to two nondimensional factors: the moisture scale height, which controls precipitation amounts, and a delay time, which controls precipitation patterns (see Barstad and Smith, 2005). We focus on the impact of changes in the nondimensional delay time as these effectively represent the dominant precipitation phase. Although the effect of precipitation amounts on landscape evolution has generated considerable study (e.g., Bonnet and Crave, 2003), the importance of the spatial distribution of precipitation in a dynamically coupled system has received little attention. Willett (1999) and Beaumont et al. (2001) specified spatial patterns in precipitation that are constant in time, to explore the interactions between erosion and tectonic deformation. Roe et al. (2002) presented a one-dimensional coupling of orographic precipitation and fluvial incision that produces within-basin variability in precipitation. Here we examine in detail the effect of changes in the spatial pattern of precipitation on the coupled system of orographic precipitation and erosion.

The nondimensional delay time is defined as $\tau^* = \frac{U\tau}{a}$, where U is the wind speed (m/s), τ is the combined growth and advection delay time (s), and a is the mountain half-width (m). τ^* compares the distance precipitation particles are advected to the size of the range. It is the dominant control on the precipitation pattern. As τ^* increases, precipitation is advected into the center of the range and, for $\tau^* > 1$, to the lee of the range. Changes in τ^* can be interpreted as representing changes in the precipitation phase: the order of magnitude slower fall of snow compared to rain allows for significantly greater penetration of snow into the center of a given mountain range relative to rain. Faster wind speeds and smaller mountain ranges also favor precipitation penetrating the center of the range. Plausible changes in climatological wind speed are not as large as changes in the fall speed of rain versus snow; therefore, they are less likely to induce significant changes in τ^*. We do not focus on the role of range width in changing τ^*, but suggest that the drying of the interior of widening orogens may influence their evolution.

Experiments were run varying τ^* across a large range. The variation of τ^* produces moderate changes in precipitation amounts as well as precipitation pattern. These changes result in lower slopes where precipitation is increased. There are also changes in the morphology of the landscape, discussed below, that cannot be accounted for with uniform changes in precipitation amounts, but that reflect changes in pattern. Two sets of models were run. One set had a dominant wind direction perpendicular to the strike of the range. In the second set of model runs, winds came from 10 equally spaced directions with equal frequency. These models illustrate the importance of intrabasin variability in precipitation without the impact of large-scale asymmetry in the precipitation pattern.

MODEL RESULTS

At the scale of the entire mountain range, τ^* has a profound impact on topography and precipitation patterns (Fig. 2). A preferred wind direction produces a rain shadow and an asymmetric topography. The main drainage divide is displaced downwind of the center of the domain and the highest peaks are located downwind of the divide. Migration of the drainage divide away from the prevailing wind is also observed in a model of crustal deformation with imposed precipitation asymmetry (Willett, 1999). The displacement of the drainage divide reaches a maximum at moderate τ^* because there is sufficient precipitation in the headwaters of windward-side rivers to capture lee-side area, and a large-enough difference between windward- and leeward-side precipitation that the lee-side rivers cannot compete.

For small τ*, maximum precipitation occurs on the windward flanks of the range and both the crest and lee side are relatively dry. Precipitation is highly variable across the domain. The topography reaches high mean and maximum elevation, is steep in the center of the range, and has a broad plain at low elevation on the windward side. In contrast, when τ* is large, precipitation reaches a maximum in the center of the range and is less variable across the range. Despite a decrease in mean precipitation, the mean and maximum elevations are lower than with small τ*. In addition, the slope of the mean topography is lower and more uniform than for small τ*.

The distribution of precipitation along river channels is important in shaping landscapes. For large τ*, precipitation increases toward the divide in the trunk streams, which allows for shallower river slopes in the headwaters than if precipitation were uniformly distributed. For small τ*, precipitation decreases toward the divide in the trunk streams on the windward side, which forces them to maintain steep slopes in the headwaters. This is consistent with the behavior of modeled one-dimensional river channels (Roe et al., 2002). In the model framework, changes in the concavity of streams are propagated throughout the landscape. The precipitation rate in the center of the range controls mean and maximum elevation, so that for large τ* the steady-state range is lower than in a case with small τ*, despite a lower average precipitation rate.

At the scale of ridges and valleys, there is a strong relationship between precipitation and topography: ridges receive more precipitation than valleys. Elevation and precipitation residuals are strongly correlated with correlation coefficients of 0.6–0.7. Precipitation is increased on ridges relative to valleys at scales of ~10 km, similar to observed ridge-valley precipitation differences measured in the Olympic Mountains (Anders et al., 2007). Precipitation increases more strongly over ridges for small τ* than for large τ*. Therefore, the increased spatial variability in precipitation for small τ* occurs at both the range scale and the ridge-valley scale.

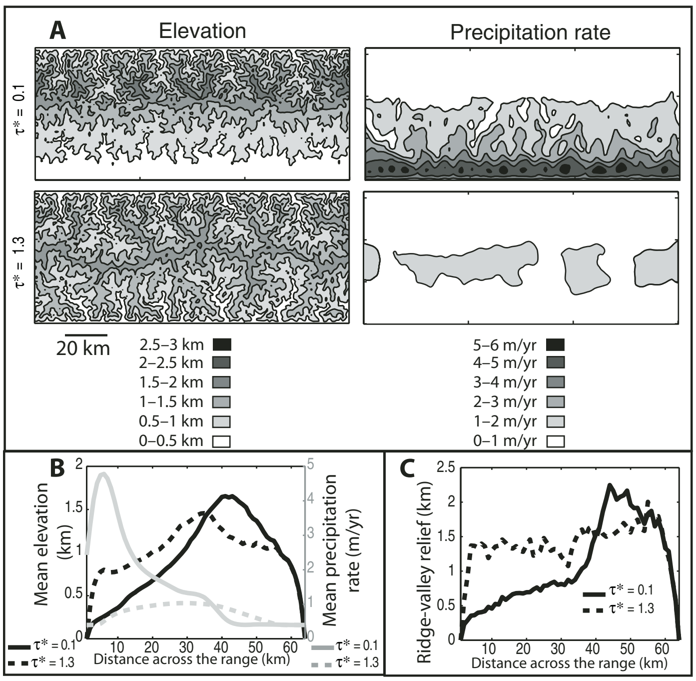

Figure 2. Coupled model results for simulations with a preferred wind direction from bottom of page in A and from left in B and C. A: Map views of elevation and precipitation rate for steady-state ranges with small (top) and large (bottom) nondimensional delay times (τ*). B: Mean elevation on left axis and mean precipitation rate on right axis as a function of distance across the domain for three values of τ*, indicated by different line styles. C: Distribution of ridge-valley relief across the range, also for three values of τ*.

The change in precipitation enhancement on ridges with τ* is accompanied by a spatially variable change in the concavity of tributary channels. For small τ*, precipitation decreases upstream in trunk streams, but increases upstream in low-elevation tributaries, leading to variability in channel concavity (a range of values from 0.43 to 0.62) both between trunk and tributary streams and with distance from the coast. Ridge-valley relief varies across the range so that, in the case of small τ*, ridge-valley relief is diminished on the flanks of the range where precipitation is strongly concentrated on ridges. As precipitation becomes more uniform across the range, ridge-valley relief and channel concavity become more uniform (Fig. 2).

Cases lacking a strong prevailing wind direction perpendicular to the strike of the range are shown in Figure 3. The ridge-valley scale relationships are similar to those described above. At the range scale, the topography lacks the asymmetry associated with a prevailing wind direction. Nevertheless, precipitation and topography show variability at the ridge-valley scale and changes in τ* produce large changes in mean and maximum elevation and ridge-valley relief.

DISCUSSION AND CONCLUSIONS

Our numerical experiments demonstrate that nondimensional delay time (τ*) is an important control on the morphology of modeled steady-state landscapes. Moreover, some of the spatial patterns of precipitation and precipitation-topography relationships are similar to measured patterns (e.g., Smith et al., 2005; Smith and Evans, 2007; Anders et al., 2007). Thus, we demonstrate the potential for observed spatial variability in precipitation to impact topography when the two equilibrate to one another.

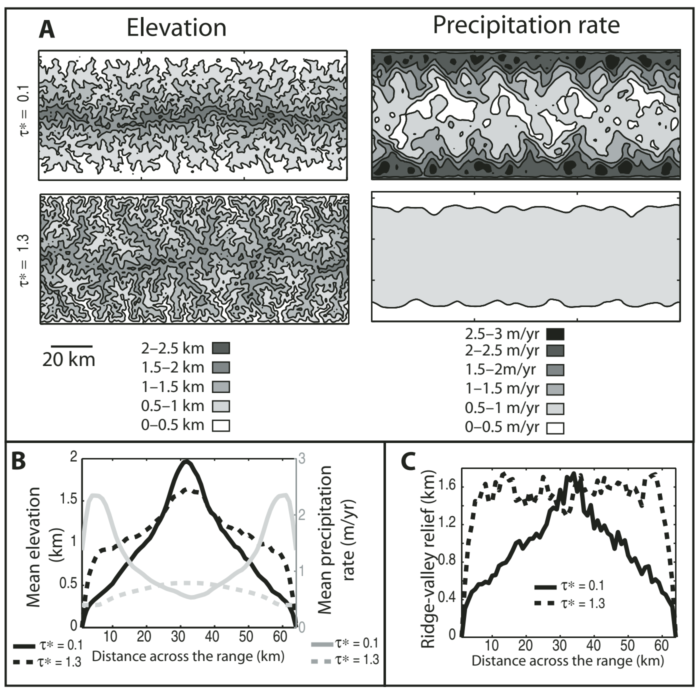

Figure 3. Steady-state topography and precipitation patterns in simulations with wind coming from 10 equally spaced directions with equal frequency. A–C as described for Figure 2.

Is there direct evidence in natural systems for the interactions between precipitation phase and landscape form observed in the model? Before asking this question, it should be stressed that both the precipitation and the landscape evolution model are first-order representations of processes known to be very complex. The idealization of climate with a characteristic wind speed and direction can produce strikingly accurate results (Smith et al., 2005; Anders et al., 2007), but may not provide a general model. The assumptions of steady, uniform, and saturated air flow are not representative of real events. The total delay time in the precipitation model represents many microphysical processes that are difficult to measure in situ and impossible to constrain over geologic time scales. Similarly, the landscape evolution model neglects the influence of spatial and temporal variability in uplift rates, variations in rock strength, and the role of sediment in mediating bedrock river incision.

Despite these challenges, the model predicts a strong response to changes in τ^*, making it more likely that a signal could be identified in natural systems. The largest difference in modeled steady-state topography comes from comparing cases with very long τ^* to those with very short τ^*. The limited data available on the delay time as estimated using long-term precipitation data are all from areas with relatively long delay times (Fig. 1). A key question is whether there are mountain ranges with short delay times over geologically significant time periods, as existing measurements in these settings are for single events. However, the strong association between precipitation patterns in individual storms and in climatological averages (e.g., Anders et al., 2007) suggests that these measurements are likely to be representative.

If the connection between warm and cool climates and average delay time works as the existing data suggest, we infer that, in the absence of strong spatial patterning in rock hardness or tectonic uplift, topography will evolve toward the modeled forms. We propose two metrics that can be evaluated within a single mountain range: the distribution of ridge-valley relief and the average slope of the topography as a function of distance across the range. These within-range measures are desirable because they provide an internal reference frame rather than relying on the absolute magnitude of slope or relief.

Specifically, we predict that rain-dominated climates will evolve toward topography with variable ridge-valley relief: relief will be lowered in wet regions on the edge of the range, while ridge-valley relief will be greater in the dry center of the range. In contrast, cool climates are predicted to be associated with nearly uniform distributions of ridge-valley relief across the mountain range. In addition, we predict that the average cross-sectional form of mountain ranges will vary, with rain-dominated ranges having an average slope that increases toward the center of the range, and snow-dominated ranges having a region with a constant slope in the center of the range.

The model predicts decreased total relief in cool climates relative to warm climates. This is accomplished entirely by fluvial processes without explicit consideration of glacial processes. It is consistent, however, with the association of Cenozoic cooling and increased erosion rates (e.g., Molnar and England, 1990). More generally, we demonstrate that the relationship between precipitation patterns and topography should vary as a function of climate. Therefore, the feedback between precipitation patterns and topography will depend on climate. This link between atmospheric and geomorphic processes emphasizes the coupled nature of the Earth system and represents an under-explored area in the interface between climate, erosion, and tectonics.

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