Coupled Evolution of Topography and Orographic Precipitation in Varied Climates

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Abstract

Precipitation patterns in mountain ranges are strongly controlled by topography, highly variable in space, and persistent over timescales for which measurements exist. The impact of spatial patterning of precipitation on landscape evolution is examined with a coupled model of orographic precipitation and surface erosion. For a range of plausible climate variables, steady-state precipitation patterns vary from nearly uniform and maximizing over the highest topography to highly spatially variable, closely coupled to small-scale topography and maximizing low on windward slopes. Such precipitation patterns are a first-order control on significant aspects of modeled landscape morphology including range scale and ridge-valley scale relief, channel concavity, and the position of the main drainage divide. An association between more uniform precipitation patterns and cooler climates is suggested based on field studies and the strong dependence of precipitation fall speed (and hence advection distance) on precipitation phase. The coupled evolution of precipitation patterns and topography further demonstrates the fundamental importance of the coupled climate, erosion and tectonic system for mountain geomorphology.

Motivation and Model Framework

It is axiomatic that precipitation exerts some control on nearly all erosional processes and thereby influences landscape evolution. Precipitation accumulated as river discharge is fundamental in driving landscape change and the geodynamic evolution of active orogens (e.g., Whipple, 2004), and glacial erosion rates likewise scale with ice discharge and thus with the precipitation rate of snow (e.g. Hallet, 1996). However, precipitation is not simply a driver of erosion and a cause of topographic change; it is strongly influenced by topography in mountainous regions. Thus the spatial distribution of precipitation and topography must evolve together. Large gradients in precipitation (factors of 2- 4) over short distances (~10-40km) are tied to topography and persist over years to decades of measurement in the European Alps (Frei and Schär, 1996), the Himalaya (Anders et al., 2006; Barros et al., 2006) and the Olympic Mountains of Washington State (Anders et al., 2007). Spatial patterning of precipitation linked to topography has clear potential to influence landscape evolution.

A coupled model combining the linear orographic precipitation model (Smith and Barstad, 2004) and the CASCADE landscape evolution model (Braun and Sambridge, 1997) was developed to explore the co-evolution of precipitation and topography. The landscape evolution model represents fluvial erosion via the unit stream power model of fluvial incision (exponent of ½ on discharge, 1 on slope) It also includes a threshold slope set at 30 degrees to represent hillslope processes. The model is run at a spatial resolution of 1km, thus, fluvial processes dominate landscape evolution. A simple fluvial incision model isolates the effects of spatially variable precipitation without additional factors such as variability in discharge or sediment in the river channel, which are neglected here although they are likely important in real systems for some time and length scales. A limited set of simulations confirm that for other common fluvial incision laws (shear stress and total stream power) results remain similar.

The linear orographic precipitation model provides an idealized representation of the relationship between precipitation and topography. The air is assumed saturated through the troposphere and has an adiabatic temperature profile. Horizontal wind speed and direction are constant in space and time. Wind speed and direction, surface temperature, and the moist static stability of the atmosphere (resistance to flow over topography) are imposed (Smith and Barstad, 2004). The flow of air over topography is then computed to determine where and how rapidly air is rising. In regions of rising air, adiabatic cooling leads to supersaturation and condensation of water vapor. A delay time represents the characteristic time scale over which water vapor is converted into precipitation hitting the ground. The delay time includes the characteristic time from the initial condensation of cloud droplets to the growth of rain drops or snow flakes large enough to have a downward directed velocity. In the atmosphere, initial nucleation of water vapor into cloud droplets of ~10 microns in diameter is followed by growth of about two orders of magnitude to reach the size of typical falling particles (e.g., Rogers and Yau, 1989). Once droplets are large enough to fall, they are still advected by the horizontal winds. The time taken for the precipitation to reach the ground depends on both the terminal velocity and the elevation at which the precipitation forms and is also included in the delay time. Different choices of delay time can simulate observed patterns of precipitation in different settings (e.g. Barstad and Smith, 2005).

The steady-state reached in this coupled system of precipitation and landscape evolution is dependent on two non-dimensional factors that together are the dominant controls on the precipitation pattern (Barstad and Smith, 2005). These two factors depend on external climatic conditions and also on the width of the mountain range. The first factor is the non-dimensional moisture scale height, defined as

$$\widetilde{H} = \frac{N_m H_w}{U} \qquad (1)$$

where Nm is the moist Brunt-Väisälä frequency, a measure of the resistance of the air to flow over topography, Hw is the moist layer depth, which depends on the surface temperature, and U is the wind speed. The lengths considered are the moisture scale height and the depth of penetration of the lifting generated by topography. More moisture is available if waves propagate through the entire moist layer depth than if waves do not lift the entire moist part of the atmosphere. Thus, this factor controls mainly the amount of precipitation generated. The second factor, non-dimensional delay time, is defined as

$$\widetilde{\tau} = \frac{U\tau}{a} \qquad (2)$$

where U is the wind speed, t is the delay time and a is the mountain half width. This factor compares the distance precipitation is advected downstream to the size of the mountain range. It dominantly controls the precipitation pattern. As non-dimensional delay time increases, precipitation is advected farther into the center of the mountain range and, eventually, for non-dimensional delay time greater than 1, to the lee of the range.

Anders (2005) explored the entire plausible parameter space of the coupled model in terms of these two non-dimensional parameters and concluded that both the amount of precipitation and the spatial distribution of precipitation are crucial in determining the characteristics of topography. The relationship between precipitation amounts and the form of steady-state topography has been studied previously (e.g., Bonnet and Crave, 2003; Whipple and Meade, 2006). When the uplift rate is held constant, and precipitation is increased uniformly the erosion rate required to balance uplift can be achieved with lower slopes, therefore lower slopes and lower relief are found at steady-state when compared to drier climates. The importance of the spatial distribution of precipitation in a coupled system has not been previously examined, to our knowledge, except in the case of a rain-shadow at the scale of the entire mountain range (e.g. Koons, 1989; Willett, 1999). In this paper we examine the impact of changes in the spatial pattern of precipitation on the evolving coupled system.

Model Results

To investigate the impact of changing precipitation patterns on steady-state topography the non-dimensional delay time was varied across a large range of reasonable values with non-dimensional moisture scale height held constant at an intermediate value (1.6). Model parameters fixed for all runs include the domain size (64 by 256 grid points), time step (1 year), uplift rate(2mm/yr), fluvial incision constant (6 x10-6 m-1/2yr-1/2) wind speed (10 m/s), surface temperature (280 K), moist Brunt-Väisälä frequency, (0.007/s), and the fraction of time that precipitation occurs (0.034). Delay time was varied from 200s to 2000s to produce changes in the non-dimensional delay time from 0.1 to 1.3. In addition to controlling the precipitation pattern, changes in the delay time also produce moderate changes in precipitation amounts which impact the resultant topography by adding a tendency for lower slopes with increased precipitation, as found in previous studies. However, the precipitation patterns produce additional dramatic changes in morphology, discussed below, that cannot be accounted for with uniform changes in precipitation amounts.

All simulations were run until a complete steady-state was reached. To facilitate comparison of topographies, one set of model runs was designed to maintain a constant mountain half-width. In these runs, winds come from ten equally spaced directions centered to be perpendicular to the long side of the model domain. The 10 resulting precipitation patterns are equally weighted to produce a composite precipitation pattern. These simulations result in ranges without rain shadows, but still create spatial variability in precipitation that is tied to topography. Changes in the non-dimensional delay time produce significant changes in both precipitation patterns and topography. Two spatial scales are examined in greater detail: that of the entire mountain range (~60 km) and that of individual drainage basins (~10 km).

At the largest scale, the non-dimensional delay time has a profound impact on topography and precipitation patterns (Figure 1). For small non-dimensional delay times, maximum precipitation is centered on the flanks of the mountain range and the center is relatively dry. Precipitation is highly variable across the domain. The resulting topography reaches high mean and maximum elevation, rises steeply toward center of the range, and has broad plains at lower elevations. In contrast, when the non-dimensional delay time is long, 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 elevation are lower than in a case of short non-dimensional delay time. The slope of the mean topography is lower and more uniform than in the case of short delay time.

These topographic changes highlight the importance of the distribution of precipitation along river channels in shaping landscapes. At long delay times, precipitation increases toward the drainage divide in trunk streams, which allows river slopes to increase more slowly than if precipitation were uniform. At short delay times, precipitation decreases toward the divide in trunk streams forcing them to steepen more quickly than in the case of uniform precipitation. This is consistent with the behavior of modeled one-dimensional river channels under variable precipitation (Roe et al., 2002). In the model, such changes in the concavity of streams are propagated throughout the landscape. The amount of precipitation delivered to the center of the range controls mean and maximum elevation, such that for long non-dimensional delay times it is possible for the steady-state range to be lower than in a case with short delay times despite lower precipitation on average in the long-delay time case.

In addition to the cases described above which have no prevailing wind direction, cases with a strong prevailing wind direction perpendicular to the strike of the range were examined (Figure 2). At the range scale, a preferred wind direction produces a rain shadow and an asymmetric topography. The main drainage divide is displaced downwind and the highest peaks are located even farther downwind than the divide (Figure 2). These features all vary with non-dimensional delay time. For short non-dimensional delay time the precipitation rate is very high on the lower windward flank of the range and mean and maximum elevation in the range interior are high. As delay time increases, precipitation on the lee side increases and precipitation on the windward side decreases, making the rain shadow less pronounced. The degree of displacement of the drainage divide reaches a maximum at moderate non-dimensional delay time as there is sufficient precipitation in the headwaters of windward-side rivers to allow them to capture lee-side area and a large enough difference between windward and leeward side precipitation so that the lee-side rivers cannot compete.

Interpretation of Results and Implications

The coupled model of orographic precipitation and landscape evolution demonstrates that the feedback between precipitation and topography is a major control on the modeled steady-state topography at the scale of the entire range and at the scale of individual ridges and valleys. Spatial patterning of precipitation is a first-order factor in determining the morphology of these modeled landscapes. The spatial patterns of precipitation produced are reasonable when compared to long term precipitation patterns measured in mountain ranges (e.g., Smith et al., 2005, Smith and Evans, 2007; Anders et al., 2007). Thus, the model demonstrates the potential for observed spatial variability in precipitation to have a large impact on topography when the two equilibrate with one another. Spatial variability in precipitation is a primary feature of mountain climates and in has been shown to be persistent over timescales from years to decades in several ranges (e.g., Frei and

Schär, 1996; Anders et al., 2006; 2007) Is it reasonable to expect a geomorphic signature of the feedback between precipitation and topography may be present in real topography?

We recognize that the idealized coupled model and the assumptions implicit in it make direct interpretation of the results in terms of natural systems difficult. The parameters in the precipitation model should not be interpreted too rigidly when considering natural climatic systems. The spatial pattern of precipitation in the coupled model is controlled by the non-dimensional delay time which compares the characteristic distance precipitation is advected with the width of the mountain range (equation 2). If range width increases while the advection distance of precipitation remains the same, precipitation tends to fall out before reaching the center of the range. Precipitation decreases in the center of the range, maximum elevations increase, and river profiles steepen relative to a narrower orogen. The model suggests that as a range grows in width in a stable climatic setting, the efficiency of the climate at eroding the high topography decreases because less precipitation penetrates into the core of the range. This implication is consistent with the observation that the width of the Andes increases in the vicinity of the high and dry Altiplano (Montgomery et al., 2001).

The characteristic distance precipitation is advected is more difficult to interpret than range width. In the model, a characteristic wind speed and the delay time determine this distance. These quantities are difficult to measure and how they may vary between ranges or over long timescales is unknown Regional storm climatology is difficult for global climate models (CGMs) to reproduce. The position and intensity of storm tracks during the last glacial maximum as modeled by a set of GCMs varies considerably from one model to another, even in the sign of changes

(Kageyama et al., 1999). Thus, understanding how storm wind speed may vary in space and time is difficult. Likewise, the delay time represents numerous microphysical processes that are difficult to observe directly and are virtually impossible to constrain over long timescales. However, one component of the advection distance is well characterized, namely, the strong control of precipitation phase on terminal velocity of falling precipitation. Rain droplets fall at 5-10 m/s, while snow falls an order of magnitude more slowly at 0.5-1 m/s (e.g., Locatell and Hobbs, 1974; Braziersmith, 1992). The much slower fall speed of snow allows it to be advected further downstream than rain. In the midlatitudes, it is common for rain at the ground's surface to have begun as frozen precipitation. In cool climates relative to warm climates, a larger fraction of precipitation will fall as snow for a longer portion of its descent increasing the distance that precipitation is advected.

The association of decreasing mean temperature with increasing advection distance is consistent with the few comparisons of the linear orographic precipitation model to observations. The work available suggests a systematic relationship between the delay time and the mean annual temperature (Figure 3). Case studies of individual storm events 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 long-term climatological pattern of precipitation and isotopic depletion of precipitation have been used to constrain delay times for the Oregon Coast Ranges and Cascades (Smith et al., 2005), the southern Andes (Smith and Evans, 2007), and the Olympics Mountains of Washington State (Anders et al., 2007). Considered together, these studies suggest that cooler climates produce precipitation patterns consistent with longer delay times and longer advection distances than warmer climates (Figure 3).

There are caveats in the association between temperature and delay time, and especially in the prediction of advection distances, as the mechanisms behind the observed relationship are not well understood. In particular, cool and warm climates not only differ in the fraction of precipitation that falls as snow, but also in the characteristic height at which precipitation forms. In cool climates atmospheric moisture is limited to a shallower layer so precipitation has a shorter distance to fall. This will tend to decrease the delay time, partially counteracting the slower fall speed of snow, relative to rain. The moist layer depth changes by a factor of 2-3 from cool to warm climates, which is small compared to the order of magnitude difference in the fall speed of rain vs. snow, suggesting that the dominant effect is the slower fall speed of snow in a cool climate. Microphysical processes of growth are dependent on the temperature and phase of the droplets. The processes contributing to growth are difficult to constrain and are neglected in this orographic precipitation model. In particular, it is difficult to predict if a cool climate will differ significantly from a warm climate in terms of the average time for growth of precipitation particles from cloud water. In summary, more research is necessary to better understand the relationship of advection distance and climate. Current knowledge allows us to say that there is variability in the characteristic delay time from range to range and this variability is consistent with increasing delay time in regions with decreasing mean annual temperature (Figure 3). This indicates a shift toward more uniform precipitation in cooler mountain ranges with lower maximum topography and lower channel concavities relative to warmer mountain ranges. In a cooler climate the tendency for more precipitation to fall in the center of the range will to increase erosional efficiency. Although our results do not include effects of a difference between fluvial and glacial erosion, they are consistent with an increase in erosion rates during climatic cooling.

A self-consistent coupled model of orographic precipitation and landscape evolution demonstrates that feedbacks between small-scale precipitation patterns and topography have the potential to be a first order control on mountain geomorphology. In the idealized model, a range of behavior from nearly uniform spatial distributions of precipitation to precipitation patterns that are closely coupled to topographic features <10 km in scale are possible. Maximum elevation, channel concavity, and ridge-valley relief of the steady-state topography are correlated with the non-dimensional delay time. The model results can be interpreted as reflect effects of variations in climate with cooler climates associated with larger non-dimensional delay times, a tendency for more spatially uniform precipitation, lower maximum elevations and lower channel concavities relative to warmer climates. Not only is the spatial patterning of precipitation sufficient to produce large changes in topography, but the nature of the precipitation patterns is also likely to vary as a function of climate. Therefore, we expect the feedback between precipitation patterns and topography to be manifested in different ways in different climatic settings. This link between atmospheric and geomorphic processes emphasizes the coupled nature of the earth system and represents an under-explored area in the relationships between climate, tectonics and erosion.

Acknowledgements

Thanks to Jean Braun, Ronald Smith, and Idar Barstad for sharing their model codes. Thanks to Drew Stolar for advice on model set-up.

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Figure Captions

Figure 1:

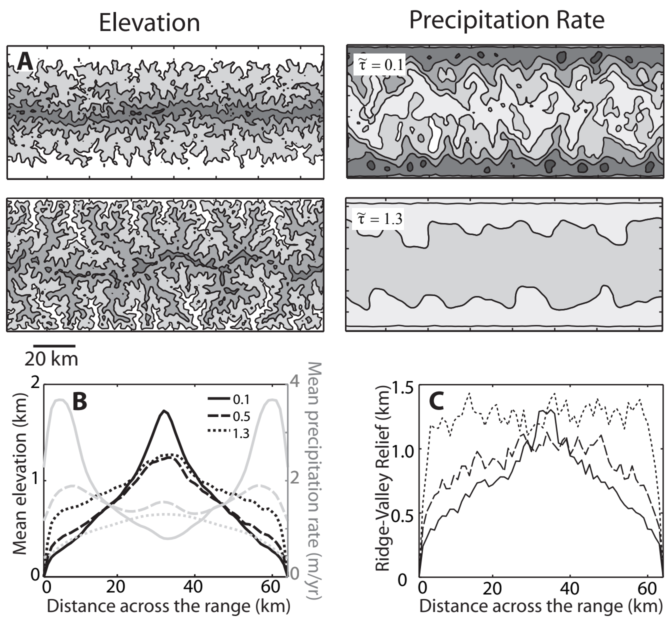
Panel A shows the map views of elevation and precipitation rate for steady-state ranges with short (0.1) and long (1.3) non-dimensional delay times. Winds come from ten equally spaced directions with equal frequency in these simulations. Elevation is contoured at 500 m intervals and precipitation rate at 1 m/yr intervals. Panel B shows the mean elevation and precipitation as a function of distance across the domain. Panel C is the distribution of ridge-valley relief across the range.

Figure 2:

Panels as in Figure 1. Results are from simulations with a preferred wind direction from the bottom in panel A and from the left in panels B and C.

Figure 3:

The estimated delay time is compared with mean annual temperature. Studies of individual events are shown as open diamond symbols and studies of long-term precipitation patterns are shown as filled squares.



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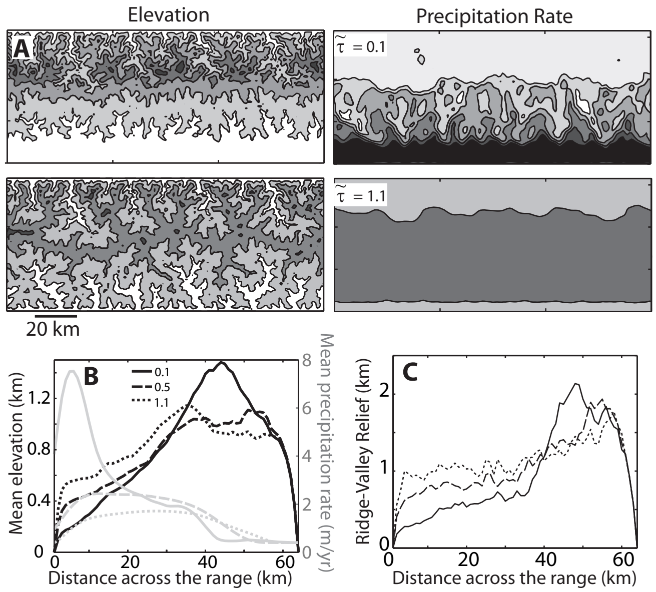

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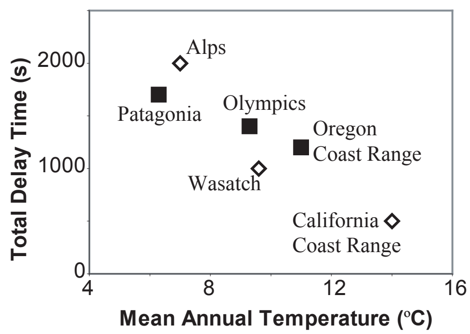

Figure 3:
The estimated total delay time is compared with mean annual temperature. Studies of individual events are shown as open diamond symbols and studies of long-term precipitation patterns are shown as filled squares.